EVOLUTION OF THE MURZUQ BASIN, SOUTHWEST LIBYA, AND SURROUNDING REGION DURING THE DEVONIAN.

Keith Robert Adamson, B.Sc.

A thesis submitted in fulfilment of the requirements for the degree of Ph.D.

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Declaration

This work has not previously been accepted in substance for any degree and is not being concurrently submitted in candidature for any degree.

Signed K. Adam ....(candidate)

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SUMMARY

During the Devonian the area currently delimited by the Murzuq Basin in SW Libya, was part of a slowly subsiding, extensive, low angle, platform cratonic margin, situated on northern Gondwana. Phases of tectonism occurred during the Gedinnian, Emsian, Frasnian and Fammenian which caused the localised uplift and erosion of regions within the basin and created areas of differential subsidence and sedimentation that persisted throughout the Devonian.

During the Early Devonian continental deposition prevailed across much of northern Gondwana, with up-dip areas such as the Murzuq Basin, covered by large, coarse-grained, siliciclastic braided and meandering alluvial systems, feeding fluvially dominated deltas and tidally influenced shorelines to the NW. During the Mid to Late Devonian marginal to open marine conditions prevailed across much of northern Gondwana, including the Murzuq Basin, with subordinate continental deposition. The up-dip alluvial systems had braided and meandering channels, passing down-dip to the NW into fluvially dominated deltas flanked by tidally influenced fine-grained shoreface successions.

During the Devonian relative sea level fluctuations led to the formation of regionally extensive sequence boundaries and flooding surfaces. The Lower Devonian succession in the Murzuq Basin contains 7 sequences that record a transgressive sequence set while the Middle to Upper Devonian succession contains a further 6 sequences formed during a general highstand. The relative sea level fluctuations that controlled the formation of these sequences were of 2^{nd} to 6^{th} order duration. Correlation of the Emsian, Givetian and Frasnian relative sea level fluctuations across North Africa and North America indicate their eustatic origin. Available data does not allow the driving mechanisms of the remaining sequences to be confidently interpreted. However, the continental to marine sequences deposited on this slowly subsiding cratonic margin were influenced by local, regional and eustatic processes.

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Chapter 1. Introduction.

This thesis is concerned with the sedimentological and structural evolution of the Murzuq Basin, situated in SW Libya, during the Devonian. The Murzuq Basin is an oval shaped topographic feature which covers about 350,000 kilometres^2 situated approximately between 10° and 15° E and 23° and 28°N (Fig. 1.1). The basin is delimited by the Tibesti and Hoggar mountainous regions in the east and west respectively, and the Gargaf plain to the north, a broad topographic uplift (Fig. 1.1). The region experiences desert climatic conditions throughout the year, with summer temperatures in excess of 50°C and an average rainfall of less than 10 millimetres (Seidl and Rohlich 1984).

This project was started in July 1995, comprising a PhD studentship at the University of Wales, Aberystwyth, undertaken in collaboration with LASMO plc, a British independent oil company with substantial interests in North Africa. This study included outcrop data that was collected during a reconnaissance field trip and a two month field season to the Murzuq Basin in 1996, and a third visit to the region in 1998 on a short conference field trip. The outcrop data obtained during these visits were correlated with subsurface data provided by LASMO plc and previously published work from the Murzuq and other basins in North Africa and around the world. Previous research relating to the pre-Devonian setting and the Devonian stratigraphy are given in the relevant chapters.

The thesis is divided into 7 chapters. Chapter 1 gives the outline of this thesis, its main objectives and provides the pre-Devonian and Devonian geological setting of the Murzuq Basin. Chapter 2 outlines the various methods used during the study. Chapter 3 contains outcrop sedimentary data and interpretations relating to Lower Devonian rocks on the SW flank of the Murzuq Basin. Chapter 4 gives details of the outcrop sedimentary data and interpretations for the Middle to Upper Devonian succession on the northern margin of the Murzuq Basin. Chapter 5 contains subsurface data and interpretations from the central area of the Murzuq Basin, which are correlated with the outcrop data and interpretations presented in chapters 3 and 4. Chapter 6 summarises data and interpretations from a number of basins in North

Africa and worldwide which are compared and correlated with the interpretations presented in chapter 5. Chapter 7 synthesises the conclusions from chapters 3, 4, 5 and 6 and discusses key points raised within them.

1.1. Thesis objectives.

This project was undertaken with the aim of furthering the understanding of the sedimentary and tectonic evolution of the Murzuq Basin during the Devonian. The stratigraphic interval was selected because it had not been studied using modern, sequence stratigraphic techniques, because it is well exposed on the flanks of the Murzuq Basin and it is also the secondary reservoir target for LASMO plc in the basin. The succession lends itself to detailed facies analysis, palaeoenvironmental interpretation and sequence stratigraphic analysis.

One of the key aims of this thesis was to contribute to the knowledge about the sedimentary and tectonic evolution of intracratonic basins that are poorly documented. The interpretations from sedimentary rocks in the Murzuq are compared to those documented from other basins in North Africa and around the world, primarily to assess the significance of the sequence stratigraphic correlations within and between these regions.

1.2. The geological setting of the Murzuq Basin.

The Murzuq Basin is one of a number of intracratonic basins in North Africa, the locations of which can be seen in figure 1.2. The present day outlines of these basins resulted from multiple phases of tectonism during the Phanerozoic (Glover 1999), and are not representative of their morphology during much of the Palaeozoic, including the Devonian when the study interval was deposited.

The present day outline of the Murzuq Basin formed during the Mesozoic (Bellini and Massa 1980). The Tibesti and Hoggar mountainous regions to the east and west of the Murzuq Basin are composed of Precambrian crystalline basement that was uplifted during the Mesozoic (Bellini and Massa 1980). The Gargaf Arch which defines the northern margin of the basin was also uplifted during post-Devonian times

(Glover 1999) (Fig. 1.3). Prior to the Mesozoic tectonic activity the Murzuq Basin did not exist as a sedimentary basin, but was part of a vast 800km wide ramp margin which stretched along the northern flank of the African part of Gondwana. Throughout the Palaeozoic this ramp margin contained a number of depressions and highs, the orientations of which were different to the subsequent Mesozoic basins and massifs.

For simplicity and convenience, the term 'Murzuq Basin' is used throughout this thesis to describe the geographical area currently within the present day basin outline shown in figure 1.3, even though it was not a separate structure during the Devonian.

To understand the mechanisms influencing the sedimentary and tectonic evolution of the Murzuq Basin and surrounding region during the Devonian it is important to consider the evolution of the region prior to this time. A detailed analysis of these data is beyond the scope of this study, but a synopsis follows, with the main stratigraphic subdivisions within the Murzuq Basin shown in figure 1.4.

1.2.1. *Precambrian to Cambrian stratigraphic evolution.*

The continental crust underlying North Africa was caught up in an early stage of the Pan-African orogeny, approximately 725 Ma ago, resulting from the collision between the Archean West African craton and the Palaeoproterozoic East African block (Lesquer et al. 1988; Dautria and Lesquer 1989; Villenueve and Cornee 1994). According to Scotese et al. (1979) this orogenic event was also responsible for the accretion of the Gondwanan supercontinent.

During the Neoproterozoic to early Cambrian the Gondwanan continent straddled the equator, stretching from mid northern latitudes to high southern latitudes, with the South Pole positioned near the western limit of North Africa (Scotese and Barrett 1990; Scotese and McKerrow 1990) (Fig. 1.5). The oldest sedimentary rocks within the Murzuq Basin are found in the Mourizidie Formation which is broadly dated Upper Precambrian to Eocambrian by Bellini and Massa (1980), with Upper Proterozoic sedimentary rocks recorded in the north of Libya

(Khoja et al. 1998). These Precambrian sedimentary rocks occur in localised areas of the Murzuq Basin, unconformably overlying the Precambrian basement in the SE of the basin (Bellini and Massa 1980). The Mourizidie Formation in the Murzuq Basin contains clasts of Precambrian schists up to 10 metres across, interpreted by Bellini and Massa (1980) as broadly coeveal with the glacial facies present further to the west in North Africa.

A rise in relative sea level during the late Precambrian to late Cambrian led to the transgression of the Murzuq Basin and surrounding region, depositing the Hassawnah Formation, interpreted by Khoja et al. (1998) as fluvial, deltaic, inter-tidal and offshore bar coarse-grained clastic deposits. These sedimentary rocks are recognised by Bellini and Massa (1980) and Khoja et al. (1998) to overlie unconformably the Precambrian basement across much of the basin, the lower contact being an expression of the peneplanation of the craton (Dautria and Lesquer 1989; Khoja et al. 1998).

A phase of Cambrian tectonism is also recognised within the Murzuq Basin by Klitzsch (1981) and Semtner and Klitzsch (1994), resulting in the generation of a NW to NNW striking structural trend that can be seen in figure 1.6.

1.2.2. *Ordovician stratigraphic evolution.*

According to Scotese et al. (1990) by the early Ordovician Gondwana had drifted slowly southwards with the South Pole still situated at, or near, the western limit of North Africa. The Murzuq Basin was positioned in high southern latitudes (Fig. 1.5). Within the Murzuq Basin the upper Cambrian Hassawnah Formation is conformably overlain by coarse-grained, cross-bedded siliciclastic facies of the Ash Shabiyat or Haouaz formations, the lower units of which may be of late Cambrian age. The type sections of these formations are defined within different areas of the Murzuq Basin and are interpreted by Khoja et al. (1998) to be coeval with one another. On the SW flank of the basin the Ash Shabiyat Formation comprises an interbedded succession of cross-bedded continental and marine sandstones, the latter units containing an ichnofabric dominated by abundant *Tigillites* and *Skolithos* traces.

By the late Ordovician, Gondwana had drifted SW, with the South Pole situated in western North Africa or the northern part of South America according to Scotese and Barrett (1990) and Scotese and McKerrow (1990) (Fig. 1.5). During the early part of the late Ordovician a rise in relative sea level resulted in the deposition of the shales and siltstones of the Melaz Shuqran Formation above the sandstones of the Ash Shabiyat Formation. This phase of deposition is interpreted by Khoja et al. (1998) to be coeval with the glaciation of western North Africa and the northern part of South America, with the Murzuq Basin situated within a periglacial shelf sea. The Melaz Shuqran Formation is typically overlain by the Mamuniyat Formation, the latter formation comprising a succession of fluvial, deltaic and shallow marine sandstones. Examples of the Melez Shuqran shales and siltstones are noted throughout the Murzuq Basin below, and also within, the sandy Mamuniyat Formation. This relationship illustrates that the two formations are diachronous with the Melez Shuqran Formation representing fine-grained, low energy open marine/basinal deposits and the Mamuniyat Formation comprising coarse-grained, high energy marginal marine to continental deposits. The interfingering of the two formations is interpreted to illustrate repeated periods when relative sea level fluctuated and facies belts migrated up-dip, or basinward.

On the SW flank of the Murzuq Basin a number of possible dropstones were observed during field studies within the Melez Shuqran Formation which are interpreted to indicate a glacial influence, corresponding to the observations and interpretations made by McDougall and Martin (1998) in the same area of the basin. Although the contact between the Melez Shuqran and Mamuniyat formations is often gradational, the former fine-grained sedimentary rocks are often eroded or absent, with lithological variations observed in closely spaced wells in the central area of the basin attributed to localised facies variations (C. Blanpied 1998, pers. comm.). Detailed analyses of seismic data from the present day central area of the Murzuq Basin by Smart (1998) and Whittington and Walker (1998) has also identified two levels of incision within the Mamuniyat Formation which are characterised by steep sided anastomosing channels that they interpret as sub-glacial ice tunnel valleys.

Similar geomorphic features are reported by Bennacef et al. (1971) within the broadly coeval facies that outcrop on the southern flank of the adjacent Illizi Basin (Fig. 1.2). These interpretations indicate that the Murzuq Basin was subject to at least two periods of ice sheet cover, with the absence of other glacially derived sedimentary features being explained by subsequent erosion (Deynoux 1998). Glacially derived and influenced deposits are also recognised in Upper Ordovician sedimentary rocks in other North African basins as well as in Arabia, Spain and France (Beuf et al. 1971; Scotese and Barrett 1990; Blanpied et al. 1998; Vaslet 1998).

The uppermost units of the Mamuniyat Formation on the SW flank of the Murzuq Basin contain evidence of increasing marine influence, while subsurface data from the central area of the basin show increasing amounts of fine-grained material and local shale deposition.

The NW-and NNW-striking structures formed during the Cambrian were also active during the mid to late Ordovician according to Klitzsch (1981), resulting in localised variations in the thickness of Ordovician deposits within the Murzuq Basin. This phase of tectonism formed a series of NW striking uplifts and troughs that persisted into the early Silurian (Klitzsch 1981).

1.2.3. *Silurian stratigraphic evolution.*

The early Silurian was characterised by a regionally extensive rise in relative sea level that began during the uppermost Ordovician. The mechanisms driving this transgressive event are difficult to constrain, although Van Houten and Hargraves (1987) pointed out that the northward drift of Gondwana during the late Ordovician to early Silurian (Fig. 1.5) carried the continent away from the South Pole, resulting in continental de-glaciation and glacioeustatic sea level rise. This transgression reached as far south as present day 15°N in North Africa, submerging the structural highs within the Murzuq Basin which had been uplifted and eroded at the end of the Ordovician (Klitzsch 1981; Semtner and Klitzsch 1994). This submergence led to the deposition of the shale- dominated Tanezzuft Formation (Fig. 1.4), the basal shales of which are often highly radioactive, a signature that can be picked on gamma ray well data from localised areas of the Murzuq and other basins in North Africa. These basal shales represent an abrupt transition from primarily coarse clastic deposits during the Cambro/Ordovician to finer-grained deposits during the early Silurian.

The gamma ray 'hot' shales are overlain by a thick succession of graptolitic shales containing subordinate siltstones and fine sandstones which make up the bulk of the Tanezzuft Formation (Fig. 1.3). The siltstones and fine sandstones within the Tanezzuft Formation contain evidence of deposition above the storm wave base (Fig. 1.7). Within the Murzuq Basin the shales of the Tanezzuft Formation are gradationally overlain by shallow marine to continental siliciclastic units of the Akakus Formation (Fig. 1.8), which represents the up-dip equivalent of the fine-grained Tanezzuft Formation. The Tanezzuft and Akakus formations represent lithostratigraphic subdivisions of a coeval stratigraphic unit, with the basal sandstone marker bed at the boundary between the Tanezzuft and Akakus formations of Radulovic (1984a; 1984b) being diachronous throughout the North African basins (Fig. 1.9).

The facies belts in the Tanezzuft and Akakus formations migrated up or down the northern margin of Gondwana as the rate of relative sea level rise or fall varied. A number of high frequency, low magnitude relative sea level fluctuations can be identified within the Silurian succession in Libya, as outlined by Bellini and Massa (1980) and shown in figure 1.9, forming several transgressive/ regressive shale to sandstone sequences. These relative sea level fluctuations occurred within a period of overall relative sea level fall that resulted in the formation of a progradational sequence set, prograding from southern Libya during the mid Silurian to northern Tunisia by the uppermost Silurian (Bellini and Massa 1980) (Fig. 1.9). This period of relative sea level fall is recognised within other areas of Gondwana and Laurussia by many authors including Van Houten and Hargraves (1987) and is therefore of eustatic origin.

During the late Silurian Gondwana drifted northwards with the South Pole positioned in central South America or North Africa (Crowell 1981; Boucot 1985; Livermore et al. 1985; Tarling 1985; Scotese and Barrett 1990; Scotese and

McKerrow 1990; Van Houten and Hou Hong-Fei 1990), or south of South America (Van Houten and Hargraves 1987; Kent and Van Der Voo 1990). Thus the Murzuq Basin was situated in low or high latitudes depending upon the palaeogeographic reconstruction.

During the late Silurian a phase of tectonic activity is recognised by Bellini and Massa (1980) and Massa (1988) to have occurred in the Murzuq Basin. This event resulted in the folding of the Akakus Formation on the western flank of the basin, the complete erosion of Silurian sedimentary rocks from the southern part of the Tripoli-Tibesti uplift, and uplift and erosion in a number of other areas of the basin and across much of North Africa. The late Silurian tectonism is generally accepted to be associated with late Caledonian epeirogenic movements (Bellini and Massa 1980; Khoja et al. 1998). This phase of tectonism increased the effects of the relative sea level fall already underway within the basin, as well as uplifting many of the NW striking structural elements shown in figure 1.6 that were active earlier in the Palaeozoic.

1.3. Plate tectonic and palaeogeographical setting of the Murzuq Basin and surrounding region during the Devonian.

Palaeogeographical and plate tectonic reconstructions of Gondwana and other continental elements for the Palaeozoic, including the Devonian have been summarised by a number of authors including Scotese et al. (1979) and Scotese and McKerrow (1990). Similar studies by Crowell (1981), Livermore et al. (1985), Tarling (1985), Van Houten and Hargraves (1987), Kent and Van der Voo (1990) utilised palaeomagnetic data. Alternate palaeogeographic reconstructions have used the distribution of flora and fauna (Johnson 1979; Poncet 1990; Burrett et al. 1990) or climatically sensitive lithofacies (Heckel and Witzke 1979; Scotese and Barrett 1990; Van Houten and Hong-Fei 1990). These authors broadly agree that by the Early Devonian Gondwana had drifted northwards, with the South Pole positioned at or near the southern regions of South America or Africa (Fig. 1.10). By the Late Devonian the southerly drift of Gondwana moved the South Pole into the southern part of Africa or South America (Fig. 1.11). These palaeogeographical reconstructions position the Murzuq Basin in high to mid southern latitudes during the Early Devonian drifting into mid to low southern latitudes during the Mid to Late Devonian. These palaeogeographic reconstructions suggest that closure of the Iapetus Ocean between Gondwana and Euramerica occurred between the mid Silurian and Mid Devonian (Scotese 1979; Scotese and McKerrow 1990).

1.4. Key features of the Palaeozoic Basins in North Africa.

The sedimentary and tectonic evolution of the Murzuq Basin and surrounding region outlined above was strongly influenced by the overall physiographic setting of the intracratonic basins and neighbouring regions of Gondwana. The response of depositional systems to changes in relative sea level or sediment supply processes, differ between basins with different basement rheology, basin floor profiles and climatic settings. Such key features of the Murzuq Basin are discussed below.

1.4.1 *Basin geometry and structure.*

As outlined in section 1.2 the North African part of the Gondwanan supercontinent formed during the late Precambrian, with the area currently represented by the North African cratonic basins situated on its margin (Fig. 1.2). Palaeocurrent data and sedimentological analysis of Palaeozoic sedimentary rocks in North Africa have highlighted a general N or NW inclined continental margin (Klitzsch 1981; Beuf et al. 1971; Clarke-Lowes 1985). Although tectonism during the Palaeozoic resulted in the uplift or subsidence of a number of areas of North Africa, these are only believed to have locally modified the continental margin (Dubois et al. 1969), the floor of which dipped at a very low angle (Fig. 1.12).

Subsidence rates in North Africa during the Palaeozoic are believed to have been low, with just over 1000 metres of sediment deposited in the Murzuq Basin area according to Bellini and Massa (1980) over the Cambrian-Silurian stratigraphic interval (over 160 Myr). This gives an accumulation rate of 0.0065 metres/kyr, approximately equivalent to the mean accumulation rate for cratonic basins determined by Schwab (1976) and an order of magnitude less than the mean accumulation rates given by the same author within foreland basins (0.186 metres/kyr). The numerous unconformity surfaces identified within the Palaeozoic succession in Libya also represented a proportion of the 160 Myr interval but even then, this rate of deposition is still moderate to low.

The geomorphological features of the northern margin of Gondwana during the Devonian are poorly documented. However, Bellini and Massa (1980), Clarke-Lowes (1985) and Massa (1988) did not document any high mountainous regions within, or close to, the Murzuq Basin. According to Bennacef et al. (1971), Beuf et al. (1971) and Clarke-Lowes (1985) Devonian alluvial sedimentary rocks within the cratonic basins of southern Libya generally contain N or NW palaeoflow indicators. However, local deflection of palaeocurrents occurred in areas near structural features (Dubois et al. 1969). The subsequent onlap of the region to the south of the Murzuq Basin by Mesozoic sedimentary rocks has prevented the location of these sediment source areas.

The composition of the coarse-grained siliciclastic Devonian sedimentary rocks within the Murzuq Basin region and surrounding area generally resembles that of the previously deposited clastic units. Thus the Devonian sandstones are generally compositionally super-mature, composed of nearly 100% quartz (Khoja et al. 1998), the grains of which are sub-angular to rounded. This compositional supermaturity is thought to indicate the repeated recycling of sediment eroded from the already mature pre-Devonian units present to the south of the Murzuq Basin (Clarke-Lowes 1985). The phase of tectonism recorded during the Late Silurian within the study region may have produced areas of increased slope that were subjected to accelerated weathering and erosion. These uplifted areas were composed of the Precambrian to Silurian sedimentary rocks and were therefore potential sources of abundant amounts of quartz. As well as the localised areas of uplift, the pre-Devonian event may also have uplifted an area to the south of the basin (Massa 1998, pers comm.), providing a major source of pre-Devonian siliciclastic material.

1.4.2 *Key factors influencing the palaeoclimate during the Devonian.*

Palaeocontinental reconstructions (figures 1.10 and 1.11) position the Murzuq Basin in low to mid southern latitudes (15°-45°) during the Devonian. This situates the Murzuq Basin in the temperate to tropical climatic zones of a Hadley cell atmospheric circulation model (Parrish 1982) or General Cirulation Model (GCM) (Barron and Washington 1982) during the Devonian. The atmospheric circulation patterns driven by the polar to equatorial temperature gradient can be modified by the position of continents and mountain belts that also affect ocean currents (Heckel and Witzke 1979; Poncet 1990). Robertson Research International (RRI) (1989) inferred a warm arid palaeoclimate in North Africa during the Middle Devonian (Fig. 1.13), while Elzaroug and Lashhab (1998) interpreted a moist palaeoclimate during the Late Devonian. In contrast, Heckel and Witzke (1979) suggested that cool, wet temperate conditions prevailed in the Murzuq Basin during much of the Devonian. According to Scotese and Barrett (1990) the GCM only models modern climatic conditions and may not be a suitable model for ancient climates.

The interaction of the continents during the Devonian may also have had a direct effect on the palaeoclimate, causing rain shadows to form near mountain belts and influencing oceanic and atmospheric currents (Heckel and Witzke 1979). According to Scotese and Barrett (1990) the positioning of large landmasses at low latitudes can cause strongly seasonal monsoonal circulation systems, while the movement of continents into polar regions can promote glaciation in greenhouse or icehouse conditions (Worsley et al. 1984). The positive elevation of Gondwana, thought to be a function of its size (Anderson 1981; 1982), and also its positive heat anomaly inherited from Proto-Pangea (Veevers 1994), may also have caused an increase in continental area, decrease in oceanic basin volume area and therefore a lowering of sea level on the continental margins. Such changes in ocean basin geometry can affect ocean currents, which in turn affect climate.

The absence of high mountainous areas in the Murzuq Basin region during the Devonian removes another possible influence on palaeoclimate. Such topographic features can create rain shadows leeward of the prevailing winds, or conversely act as major catchments areas feeding down-dip alluvial systems (Scotese and Barrett 1990). The evolution of land plants during the Devonian may also have affected palaeoclimate through changing evapo-transpiration processes, soil properties, and fluvial hydrology and channel style (Schumm 1968; Miall 1996).

Chapter 2. Methodology

In the course of this research a number of methods were used to collate, collect and interpret a variety of data sets. These methods are outlined in this chapter. The data and interpretations are described in chapters 3 to 6 and summarised in chapter 7.

2.3. Field studies

During two field seasons in 1996 field data were collected from the Lower to Upper Devonian sedimentary rocks exposed on the western and northern margin of the Murzuq Basin (Fig 1.2). The Lower to Upper Devonian succession studied is predominantly siliciclastic and is described in chapters 3 and 4. The first visit to the Murzuq Basin in March 1996 was primarily for reconnaissance purposes to delimit project objectives and identify areas for detailed study. That visit also included a general study of the Lower and Upper Palaeozoic sedimentary rocks that crop out along the western and northern margins of the Murzuq Basin (Fig. 1.2). Some of these data and their interpretations are given in chapters 1, 3 and 4. The main field season to the Murzuq Basin was between October and December 1996, comprising 28 days of field study, mainly of the Lower to Upper Devonian succession. The Upper Devonian succession was not studied as thoroughly as the Lower to Middle Devonian due to limited time in the field. Problems were also encountered accessing the Upper Devonian exposures east of Tarut due to the prevailing political situation. The field data were collected and interpreted using the methods described in the following sections.

2.1.1. *Logged sections.*

A number of sections were logged through the Lower Devonian Tadrart and Ouan Kasa formations that crop out on the western flank of the Murzuk Basin. These sections are of varying height but total over 165 metres of vertical succession; the data collected are displayed and interpreted in chapter 3. Logged sections were also undertaken through the Middle to Upper Devonian succession studied on the northern margin of the Murzuk Basin. These logs vary in their stratigraphic coverage and detail but total over 370 metres of vertical succession through the Middle to Upper Devonian succession; these data are given in chapter 4. As well as the Devonian succession, the sedimentary rocks that immediately overlie and underlie the Devonian succession were also studied at a number of localities, and are mentioned in chapters 1, 3 and 4 when relevant. These included the Hassawnah Formation (Cambrian), the Ash Shabiyat and Memuniyat formations (Ordovician), the Tanezzuft and Akakus formations (Silurian), and also the Ashkidah Formation (Devonian/Carboniferous).

The data collected from the logged sections included the lithology, grain-size, grain-sorting, grain-shape, cement, sedimentary structures (section 2.1.3), bedform geometry, ichnofabric, and any floral or faunal content. Each logged section has been given a locality number which is referred to in the relevant text. The locality numbers for the Lower Devonian logged sections are listed in appendix 3.1 while those for the Middle to Upper Devonian are listed in appendices 4.1, 4.2 and 4.3. Where non-Devonian sedimentary rocks were logged they are indicated on the logged section and are described when relevant in chapters 3 and 4.

2.1.2. *Study areas.*

As well as the logged sections, many areas were visited where the Lower to Upper Devonian succession was studied in detail but not logged due to poor exposure and time limitations. The dataset from these areas is of the same type as the logged sections, and is integrated with data from logged sections in chapters 3 and 4. Data from several of the detailed study areas were recorded on dictaphone before transferral to a note book. As with the logged section, the detailed study areas have been given locality numbers that are listed in appendices 3.1 and 4.1-4.3 for the Lower and Middle to Upper Devonian sedimentary rocks respectively. Where non-Devonian strata were studied the data are incorporated and described when relevant in chapters 3 and 4.

2.1.3. *Palaeocurrent data.*

A variety of palaeocurrent data were collected from Lower to Upper Devonian sedimentary rocks (chapters 3 and 4), as well as a lesser amount from Cambrian, Ordovician and Silurian sedimentary rocks. The collection of representative palaeocurrent data from Middle to Upper Devonian sedimentary rocks was complicated by the nature of outcrops which are generally poorly exposed cliff sections that face northward. These outcrop patterns can lead to the collection of a biased palaeocurrent data-set from sedimentary structures and foresets oriented perpendicular to the outcrop. Special care was taken to collect data from a number of variously oriented outcrops. The data from the planar and trough bedforms were collected in terms of the amount and direction of dip of the foreset (Fig. 2.1). In addition, measurements from low angle planar laminae, epsilon cross bedding and various asymmetrical and symmetrical ripple forms were collected. The regional stratigraphic dip of the Lower to Upper Devonian succession on the western and northern margin of the basin is generally between 1 and 3° so no corrections for tilt are required during data treatment.

The palaeocurrent data were input into ROSE, a PC based analytical program from Rockware Inc. The resultant rose diagrams and interpretation of the palaeocurrent data are given in chapters 3 and 4.

2.1.4. *Photo analysis.*

Over 300 photographs were taken in the course of field studies. The photos comprise montages of cliff sections and bedform geometries as well as close-ups of sedimentary structures, ichnofabrics, and the floral and faunal content of the sedimentary rocks. The montages were used to assist the analysis of the large scale inter-relationships of sand bodies, while the photographs of sedimentary and biogenic structures, and floral and faunal specimens were taken to aid their identification on return to the U.K. Help from the following academics was also sought to aid the identification of certain sedimentary and biogenic structures;

Specialist	Institution	Field of expertise
Dr. D. McIlroy	University of Liverpool, U.K.	Palaeozoic ichnofacies
Dr J. Howell	University of Liverpool, U.K.	Sedimentology
Dr J. Howell	University of Liverpool, U.K.	Ichnofabrics

2.2. Sample analysis.

To aid the sedimentological interpretation of the Lower to Upper Devonian sedimentary rocks on the western and northern margins of the basin a number of samples were collected for petrographic and ichnofacies analysis. A small amount of sidewall core data from well A1-NC174 in the central part of NC-174 in the Murzuq Basin was also available for interpretation (chapter 5) as well as core reports from the Lower Devonian succession in southern Algeria (chapter 6). The samples collected on the western and northern margins of the Murzuq Basin were utilised for detailed hand specimen study, and for thin section petrographic analysis (chapters 3 and 4).

Samples of the various ichnofabrics, and the flora and fauna of the Lower to Upper Devonian succession were collected from the western and northern margins of the Murzuq Basin. These samples were interpreted where possible and were also sent when appropriate to the following experts in the fields of palaeontology and ichnofacies analysis.

Specialist	Institution	Field of expertise
Dr. J. Maisey	Natural History Museum, New York, U.S.A	Palaeozoic Agnathans and Ctenacanthus
Professor M. Benton	Bristol University, U.K.	Palaeozoic Ctenacanthus

2.3. Subsurface data analysis.

Various subsurface data were utilised during the course of this research, to aid interpretation of the evolution of the subsurface areas of the Murzuq Basin and the surrounding region during the Devonian (chapters 5 and 6). These data were used in isolation and in conjunction with other data sources outlined in this chapter to provide a linkage with the geographically isolated outcrop data collected on the western and northern margins of the basin.

2.3.1. *Wireline log data.*

The wireline log data from the Murzuq Basin and its surrounding basins were studied to constrain the evolution of the region during the Devonian, integrated with outcrop data to interpret the pre-and syn-Devonian tectonic and sedimentary setting of the region. The wireline log data sets from the Murzuq Basin were obtained during petroleum exploration in the 1950's by companies such as ESSO, and more recently in the 1980's and 1990's by BRASPETRO, LASMO, REPSOL, ROMPRETROL and their partners in concessions NC101, NC115 and NC174 (chapters 1 and 5). The log data set from the early exploration comprises data from the following well log types:

1) Composite lithology logs.

The original data sources for the logs are unspecified and may incorporate data from any of the following log types (2- 10). The interpretation may also include the lithological transitions picked by the analysis of chippings in the drilling mud by the mud logger.

2) Drilling time plots.

This information provides a crude estimate of lithology, short drilling time = shale, longer drilling time = sandstones. It does not take account of the effects of various types of cement and amounts of compaction on the drilling time.

3) Mud/Gas analysis plots.

The varying composition of the mud/gas can be used to interpret the lithologies being drilled.

4) Hydrocarbons in mud plots: see 3).

5) Chlorite in drilling mud plots: see 3).

As outlined by Asquith and Gibson (1982), problems can occur in using the drilling mud and its composite elements (Chlorite ppm, hydrocarbon ppm, chippings) to interpret lithological transitions in the well. The drilling mud and elements within the mud can circulate within the well column for some time and can contaminate readings made in-situ or at the surface.

The hydrocarbon exploration undertaken in the Murzuq Basin and surrounding region in more recent years has utilised wireline logging tools. Well log data from the following wireline tools were interpreted:

6) Spontaneous potential logs (SP).

SP logs measure the difference in electrical potential between the formation tested and the surface (Asquith and Gibson 1982; Emery and Myers 1996). These data are sensitive to changes in permeability and are generally used to distinguish between impermeable shales and permeable sandstones. SP readings can be used to interpret hydrocarbons, cemented horizons/zones and changes in the formation water salinity (Asquith and Gibson 1982; Emery and Myers 1996).

7) Resistivity logs.

The resistivity logging suite measure the bulk resistivity of the rock which is a function of lithology, cement, porosity and pore fluid. A porous lithology with a conductive pore fluid has a low resistivity, while a non-porous lithology (or one with a pore fluid of low conductivity such as hydrocarbon) has a high resistivity. From these resistivity data the hydrocarbon component of pore fluid can be determined (Asquith and Gibson 1982).

8) Compensated Neutron Log (CNL) and Formation Density Log (FDC).

These logging tools are the best record of lithology and can thus be used to interpret depositional trends (Emery and Myers 1996). The neutron log (CNL) measures the interaction between neutrons emitted from the tool and hydrogen within the formation which can be used to estimate formation porosity (Asquith and Gibson 1982). The density log measures the true bulk density of the formation by measuring the backscatter of gamma rays which gives the electron density of the formation. The density log data can be used to assist the identification of evaporite-rich zones, gas-bearing zones, lithology, and hydrocarbon density according to Asquith and Gibson (1982). The neutron and density logging traces are often overlaid and the horizontal separation or cross-over of the traces can be used to determine grain-size changes, coals and cemented zones (Emery and Myers 1996) which are often unresolved using other logging tools.

9) Sonic logs.

Sonic logs measure the interval transit time of a seismic pulse through the formation which is related to porosity and lithology. A shale has a lower transit time (higher velocity) than a sandstone of a similar porosity (Emery and Myers 1996). High concentrations of organic matter such as coals and black shales will result in low velocities and associated very long transit times (Emery and Myers 1996). Sonic log data are affected by post depositional cementation and compaction and as a result can be of limited use in distinguishing between sandstones and shales.

10) Gamma ray logs.

The primary wireline tool used in this research were Gamma ray logs. Gamma ray wireline logging tools measure the natural radioactivity of rocks which is a function of the amount of naturally radioactive minerals present within the rocks. The radioactive elements and minerals are:

i) Potassium (K): Present in K-feldspars, micas and illitic clays (Davies and Elliott 1996 and references therein).
ii) Thorium (Th): Concentrated in sand-and silt-sized detrital minerals present within sediments such as the monazite and zircon groups, and in the fine-grained

fraction in association with certain clay minerals and as authigenic phosphates (Davies and Elliott 1996, and references contained therein)

iii) Uranium (U): Can occur within the detrital minerals suite but can also be found concentrated in anoxic sediments.

Gamma logs record the total number of radioactive counts during a specified time period, usually counts per second. The shape of the gamma log wireline trace is often used to infer changes in grain-size and thus interpret depositional facies. However, as pointed out by Rider (1990) and North and Boering (1999), these relationships cannot be universally applied. Spectral gamma log (NGS) plots measure the concentrations and the ratios of the three most common radioactive elements (K, Th, and U) outlined above (Hurst 1990). Spectral gamma ray data are rarely used and despite recent advances in interpretation techniques (Myers and Bristow 1989; Davies and Elliott 1996), doubts still exist over the limitations of spectral gamma data without accurate log calibration (Humphries and Lott 1990). The correct evaluation, correlation and interpretation of spectral gamma data requires the 'user' to have an accurate understanding of the mineralogy of the stratigraphic succession. The generally high concentrations of the naturally radioactive elements in fine-grained sediments mean that increasing radioactivity is often associated with a higher clay content and decreasing depositional energy (Emery and Myers 1996). However, a clean sandstone with a low clay mineral content may contain K feldspars, and possibly U or Th, which result in a high gamma ray reading which can only be resolved from spectral gamma data (Asquith and Bristow 1982; Hurst 1990). A limited amount of spectral gamma ray data from well A1-NC174 in the central Murzuq Basin, was interpreted (chapter 5).

When downhole gamma ray measurements are taken the natural radioactivity of the succession within 0.3 metres of the detector is measured, usually every 0.15 metres, to provide a moving average. Recent advances in logging tool technology have helped correct the problems with gamma ray data which result from drilling mud contamination, variable logging speed, and poor repeatability of open hole testing

(Hurst 1990; Davies and Elliott 1996; Emery and Myers 1996). The total count and spectral gamma data are used to interpret lithology and effect correlation in the subsurface, in recent times using sequence stratigraphic methods (section 2.4). Gamma ray data were utilised during this research as they provided the most readily available suite of well data to correlate accurately throughout the region.

2.3.2. *Seismic data.*

A large amount of seismic reflection data has been acquired from the Murzuq Basin and surrounding basins since the 1950's. The majority of seismic data utilised here comprises part of a 2D survey acquired between 1992-1995 for LASMO plc, although some data acquired during surveys in the late 1980's were also used.

Detailed descriptions of the methods involved in the acquisition of seismic reflection data can be found in Badley (1985) and Sheriff (1985). A number of areas of interest and specific problems unique to seismic interpretation in the Murzuq and analogous basins were identified during this research and are now discussed in more detail.

2.3.3. *Seismic data; acquisition and processing characteristics.*

Most of the seismic reflection data used in this research were acquired and processed in a 2D survey of concession NC-174 (Fig. 1.2) in 1992 by Seismograph Services Limited and Geco-Prakla. The NC-174 survey (chapter 5) utilised vibroseis sources, and a similar source is assumed for the seismic data from concessions NC-101 and NC-115 (Fig. 1.2). An example of the seismic data from the Murzuq Basin can be seen in figure 2.2.

The seismic data obtained within the study region (Fig. 1.2), and utilised here (chapter 5), were industry standard processed (including migration) to provide the best image of the entire stratigraphy, including the primary hydrocarbon target stratigraphy (Silurian and Cambro-Ordovician).

2.3.4. *Problems encountered with seismic data.*

i) Murzuq Basin geomorphology.

The variations in the Murzuq Basin surface geomorphology and terrain can complicate data processing from this region. Within concession NC115 in the Murzuq Basin, the southern part of the concession is a gravely and rocky desert while the northern part of the concession is covered by part of the Awbari Sand Sea which comprises sand dunes up to 150 metres high (Seidl and Rohlich 1985). Within the Murzuq Basin there are several wadis and escarpments that cut across the area, the most prominent being the Cretaceous escarpment which is up to many tens of metres high and results in a vertical shift of 206 milliseconds on seismic sections north of the escarpment. The irregular topography can cause the dissipation of wavelet energy near the surface and also variable near surface wavelet velocity (Badley 1985) which can result in poor quality seismic data (figure 2.3). The variable terrain can also cause the distance between shotpoints to increase that can result in high static levels and overall poor data clarity (Fig. 2.3). Where there is high ambient noise or static problems seismic data processing often includes a high frequency filter (Emery and Myers 1996) which results in an increase in the width of the Fresnel zone and a decrease in the horizontal and vertical resolution of the data (Badley 1985; Emery and Myers 1996).

ii) Murzuq Basin subsurface structure.

Bedding in the Murzuq Basin generally dips at a low angle, as observed in outcrop on the western and northern margins of the basin and on seismic sections (chapter 5). As a result the processing stream applied to one part of a seismic line to create the best resolution of the target stratigraphic interval generally applies to the entire seismic line. The faults observed on seismic sections within the basin are generally poorly imaged (chapter 5), probably due to the fault zones absorbing wavelet energy (Fig. 2.4).

2.3.5. *Devonian stratigraphic interval characteristics.*

Within much of the subsurface Murzuq Basin the Middle to Upper Devonian strata immediately overlies the Tanezzuft Formation of Silurian age. According to well data from the Murzuq Basin (chapter 5) the Devonian succession is up to 110 metres thick in the northern part of NC115 and only 40-60 metres thick in NC174 and NC101. The reduced thickness and the attenuation or processing out of the higher frequency part of the signal makes resolution of the subsurface Devonian succession throughout the basin very poor. The lower contact of the Devonian with the Tanezzuft Formation can be picked only because of the high acoustic impedance contrast between the predominantly shaly Tanezzuft Formation and the Devonian shales, siltstones and sandstones (Fig. 2.5). The gradational Devonian/Carboniferous boundary provides no such acoustic impedance contrast so is not resolved on seismic lines. The NC-174 seismic data were correlated with the aid of borehole data (chapter 5).

2.4. Sequence Stratigraphic methods.

The refinement of sequence stratigraphic concepts over the last 20 years has resulted in improved methods of interpreting sedimentary rocks within a chronostratigraphic framework. This framework can be used to predict the distribution of lithofacies during relative sea level fluctuations and also aid the correlation of successions within and between basins. To aid the interpretation of, and correlation between, outcrop and subsurface datasets, sequence stratigraphic principles are extensively applied to the Devonian sedimentary rocks within the Murzuq Basin. These sequence stratigraphic techniques also enable the correlation of the Devonian succession within the Murzuq Basin (chapters 3, 4 and 5) to be compared and correlated coeval units in other basins in Africa and throughout the world. A brief synopsis of the history and key methods used in sequence stratigraphy is outlined below, followed by details of sequence stratigraphic techniques that are specifically relevant to geological setting and facies with which this thesis is concerned.

2.4.1. ***Sequence Stratigraphy, a historical perspective.***

Sequence stratigraphy is concerned with the geometry and character of rock types deposited at a particular time and how these factors subsequently evolve through time. The identification of key surfaces which separate strata is integral in the application of sequence stratigraphy. These surfaces form at various stages on a fluctuating relative sea level curve and are detailed in appendix 2.1 as are the key terms used in sequence stratigraphy.

The initial work that developed the idea of transgressive/regressive cycles was undertaken in the 1930's and 1940's with the concept of unconformity bounded sequences proposed by Sloss et al. (1949). Those authors subdivided the North American Phanerozoic succession into six sequences that could be correlated throughout the North American cratonic basins.

The sequence stratigraphic concepts outlined by Payton (1977) were developed to explain the time-significant stratal relationships observed on seismic sections. The seismic stratigraphic concepts outlined in American Association of Petroleum Geology Memoir 26 (Payton 1977) are difficult to apply to the Devonian succession in the study region for reasons explained in chapter 5. Vail et al. (1977) outlined the concept of eustasy as a control on stratal geometries, and concentrated on the 'Exxon' sequence stratigraphic methodology whereby the main emphasis is placed on erosional surfaces. Subsequent work expanded these concepts to erect a global sea level chart (Haq et al. 1988). The next major development came with the publication of the Society for Sedimentary Geology Special Publication 42 (Wilgus et al. 1988), which detailed amongst other things the concept of accommodation space (Jervey 1988). That volume also contained the conceptual sequence stratigraphic models of Van Wagoner et al. (1988); Posamentier et al. (1988) and Posamentier and Vail (1988) which form the framework for modern 'Exxon' sequence stratigraphy. An alternative method for the sequence analysis of sedimentary facies was developed by Galloway (1989 a, b) that utilised maximum flooding surfaces as genetic sequence boundaries.

The application of sequence stratigraphy to outcrop data and the integration of these data with seismic stratigraphy was the next key step forward and is summarised in Van Wagoner et al. (1990). The concepts outlined in that work have subsequently been applied to outcrop data throughout the World. While early work was concerned with the application of sequence stratigraphy to data from passive margins, more recent work has been concerned with the application of sequence stratigraphy to more varied basin types such as active margins (Allen and Posamentier 1993), rift basins (Howell and Flint 1996) and intracratonic basins. Sequence Stratigraphic methods have also recently been applied to non-marine settings by many authors including Shanley and McCabe (1993, 1994), Miall (1996) and Quirk (1996).

This thesis concentrates on the 'Exxon' derived sequence stratigraphic methods and terminology rather than those outlined by Galloway (1989 a, b). This preference is made because it has been well documented that the timing of the maximum flooding surface, the genetic sequence boundary, is strongly influenced by the rate and character of sediment supply (Van Wagoner et al. 1990; Wehr 1993). A recent summary of sequence stratigraphic methodology can be found in Emery and Myers (1996).

2.4.2. *Basic concepts: Eustasy and composite eustatic sea level curves.*

Sea level changes may result from a number of interacting factors, the relative input of each being difficult to evaluate. In broad terms, eustatic changes in sea level result from a change in either the volume of the ocean basins or the volume of water contained within those basins. The main factor controlling the volume of ocean basins is the character of ocean spreading ridges. Fast spreading ridges emplace large quantities of hot, buoyant oceanic crust that displaces water onto the continental margins. Conversely, slow spreading ridges are cool and topographically relatively subdued, displacing less water. Changes in ocean ridge spreading rates can take place over millions to tens of millions of years (Fig. 2.6). High frequency eustatic sea level fluctuations can also occur over much shorter time scales; these are typically of glacio-eustatic origin which can result from perturbations in the Earth's orbit which modify climate (Fig. 2.6). The result is a complex interaction of cycles which when superimposed on one another form a composite eustatic curve (Fig. 2.7). When combined with different rates of subsidence these sea level changes produce parasequences, sequences and mega-sequences. The effects of eustatic sea level are generally difficult to measure because of local tectonics and consequently relative sea level is typically recorded for particular basins.

To understand a simple sequence it is useful to consider a single sinusoidal curve upon which inflection points are identified (Fig. 2.8). These are the rising (R) inflection point and the falling (F) inflection point, positioned on the curve where the rate of rise and fall respectively are at a maximum (Fig. 2.8). The resultant sequence is a combination of the rate of creation or destruction of accommodation space and the sediment supply. From conditions of low accommodation, space is rapidly created and water depth increases. As the rate of sea level rise begins to decrease, through the R inflection point, sediment supply catches up and begins to fill the accommodation space. This results in water depth decrease and progradation until all of the accommodation space at a point is filled. Once accommodation space is filled at or after the top of the curve, there is erosion of the deposited sediment as the sea level falls. If the eustatic curve returns to its pre-rise level then it is possible (hypothetically) that all of the sediments deposited as a result of the rise will be eroded as zero accommodation space has been created. However, complete erosion is unlikely and an amount of compaction or subsidence will likely cause the downward movement of sediment, providing accommodation space.

2.4.3. *Basic concepts: Subsidence and Relative Sea Level (RSL).*

The conceptual sequence stratigraphic models developed by Jervey (1988), Posamentier et al. (1988) and Posamentier and Vail (1988) outline how the depositional style, distribution and evolution of sedimentary facies can be related to eustasy. By assuming a constant rate of subsidence at a point and a sinusoidal eustatic sea level curve (Fig. 2.8), a model for relative sea level or sediment **accommodation space** can be made (Fig. 2.9). The rate at which accommodation space changes through time can control the internal character and stacking pattern of parasequences, parasequence sets and sequences. If sediment supply is assumed to be constant then a **relative sea level** curve (RSL curve) can be derived by adding subsidence at a point (k) with a sinusoidal eustatic curve in terms of time (t). The equation for relative sea level (RSL) is:

$$RSL=\sin(t) + k(t)$$

It is assumed that subsidence increases in a basinward direction and also that lateral variations in relative sea level can influence the character, timing and distribution of systems tracts (Posamentier et al. 1988). As the rate of subsidence changes, the effect on parasequence to sequence scale stacking patterns varies accordingly. Figure 2.10 shows the effect of a range of subsidence curves on the sinusoidal eustatic curve illustrated in figure 2.8. Figure 2.10 also illustrates that the timing of the R and F inflection points is independent of subsidence. The RSL curves produced from a variety of subsidence curves also controls the duration and/or occurrence of the various system tract types (Fig. 3, Wehr 1993). A low rate of subsidence can result in an early shift from retrogradational (Transgressive Systems Tract-TST) to progradational (Highstand Systems Tract-HST) deposition (Wehr 1993). This low rate of subsidence will also result in the duration of the HST being shortened and an early transition to facies deposited during the Falling Stage Systems Tract (FSST) and Lowstand Systems Tract (LST). In this situation the FSST and LST dominate, with prolonged periods when there is a basinward shift in facies and extensive erosion. In many cases where subsidence is low the preservation potential of the HST deposits is low (Wehr 1993). Where the angle of the basin slope is low, erosion during the formation of sequence boundaries would be restricted to shallow, wide incised valleys, rather than the formation of steep sided incised valleys where steeper basin slopes are present (Wood et al. 1993).

When interpreting a succession in terms of its sequence stratigraphic evolution is it important to consider what is meant by relative sea level. Recent work has outlined the problems in extending sequence stratigraphic correlations from the shallow marine to continental system (Schumm 1993; Shanley and McCabe 1993; 1994). When considering continental successions the concept of base level is introduced as "the surface above which erosion occurs and below which deposition takes place" (Sloss 1962). The interpretation of non-marine successions using sequence stratigraphic methodology is outlined in section 2.4.8.

The input of sediment compaction and isostatic feedback must also be considered when interpreting RSL curves. The amount of sediment compaction is dependent on a number of factors, including: (1) the lithology of the compacting sediment (sandstone vs shale), (2) the nature of pore fluid in the compacting sediment, and (3), the nature of the load placed on the sediment. The weight of the water column on the lithosphere also has an effect on isostatic feedback, although the amount of this loading is difficult to ascertain as water depth cannot be easily be constrained from geological data (Plint 1997b).

2.4.4. *Hierarchy of Sequence Stratigraphic elements.*

Some of the key terms outlined in appendix 2.1 form a hierachy of sequence stratigraphic elements. During each high magnitude cycle of sea level rise and fall a number of low magnitude, high frequency events can often be recognised (Fig. 2.7) that do not generally cause periods of erosion; these events create parasequences (see appendix 2.1). These parasequences are the building blocks of sequences and are best identified in shallow marine settings (Van Wagoner et al. 1990). Parasequences typically record a shoaling of water depth, and in most siliciclastic systems, a coarsening-upward of grain-size representing the progradation of a delta or shoreface into accommodation space generated by a rapid rise in relative sea level. Within the coastal plain parasequences are virtually impossible to distinguish from cycles driven by autocyclic changes (Miall 1996).

Parasequences form stacking patterns bounded by either a major marine flooding surface or a sequence boundary (Van Wagoner et al. 1988). These stacking patterns can be progradational, aggradational, or retrogradational. Parasequences sit within systems tracts. Systems tracts are defined by the stacking of parasequences within them, and represent different stages of relative sea level change and sediment supply within a sequence. With the recognition of composite relative sea level curves (Fig. 2.7) that comprise 3rd, 4th, and 5th order cycles it is possible to have a number of eustatic events capable of producing sequences. It may be the case that the third order event will contain high frequency 4th order sequences that stack in a comparable way to parasequences in simple sequences.

It is possible to distinguish between 3rd and 4th order components using three main criteria:

i) The duration of the composite element, i.e. 1-5 Myr for a 3rd order sequence, 0.1-1 Myr for a 4th order sequence.

ii) The degree of facies dislocation at a 3rd order sequence boundary will be much greater than the associated 4th order event as a result of the higher magnitude of sea level fall. Also, the 3rd order flooding surface will be of greater magnitude, duration and regional extent than the 4th order flooding event, possibly producing a thicker argillaceous succession.

iii) The stacking patterns of the higher frequency (lower magnitude) events contained within the composite sequence sets can indicate when the key surfaces within the high magnitude cycle occurs.

Third order composite sequences are most commonly identified in up-dip areas where subsidence is lower and incision at each high frequency sea level fluctuation is more marked than flooding. Further down dip into the basin, where more accommodation space is available, these high frequency events will cause parasequences to occur, rather than resulting in incision.

2.4.5. *Sequence boundaries.*

During a simplified sinusoidal eustatic cycle a sequence will develop bounded by an upper and lower sequence boundary. Two main types of sequence boundary are identified:

Type 1 sequence boundary: Interpreted to form when the rate of relative sea level fall exceeds the rate of subsidence and leads to sub-aerial incision of the shelf with the development of incised valleys and interfluves.
Type 2 sequence boundary: Interpreted to form when subsidence exceeds the rate of sea level fall, sub-aerial exposure of the shelf does not occur at any stage of the base level cycle. Periods of base level fall associated with type 2 sequence boundaries cause aggradational deposits to form immediately after progradational highstand systems tract deposits.

As the rate of subsidence within the Murzuq Basin during the Devonian is low, and the basin profile dips at a low angle and has no shelf-slope break, type 1 sequence boundaries will be the dominant type of sequence boundary formed.

2.4.6. *Seismic Stratigraphy.*

The techniques of seismic stratigraphy outlined within by AAPG memoir 26 (Payton, 1977) were developed to explain the time significant stratal relationships observed on seismic sections. Reflectors terminate at some point against other reflectors, where the reflector thins below seismic resolution or where the bedding plane terminates. The character of reflector terminations is the key to seismic interpretation and represents changes in sedimentation, non-deposition, or erosion (Vail et al. 1977).

2.4.7. *Sequence Stratigraphy in Cratonic Basins.*

The conceptual models for siliciclastic sequence stratigraphy developed by Jervey et al. (1988), Posamentier et al. (1988) and Posamentier and Vail (1988) were applied primarily to passive margin basins. Such basins have profiles containing shelf slope breaks. The geotectonic setting of the Murzuq Basin region during the Devonian within the Gondwanan supercontinent was that of a platform or ramp margin which contained no such shelf-slope break (figures 1.12 and 2.11) (section

1.2). During the Devonian the nearest oceanic crust to provide such an abrupt change in basin profile was situated several thousand kilometres from the Murzuq Basin (figures 1.10 and 1.11).

Relative sea level fluctuations influencing deposition within the North African region of Gondwana during the Devonian will have had a different effect on sequence and systems tract development than relative sea level fluctuations of a similar magnitude and frequency in passive margin basins. In this region of the Gondwanan supercontinent, when high magnitude fluctuations in relative sea level occurred during the Devonian they had the potential to shift the palaeoshoreline up-dip or down-dip, possibly hundreds or thousands of kilometres as modelled in figures 2.11 and 2.12. These high magnitude fluctuations resulted in the subaerial exposure or denudation of very large areas of the interior of the continent. During major falls in relative sea level in these cratonic regions the palaeoshoreline was shifted a long distance down-dip from the fluvial systems feeding the palaeoshoreline during the previous HST and FSST (Fig. 2.11). It took the fluvial systems a long time to regrade to the equilibrium profile of the new coastal plain due to the long distances involved. This would have led to the new detached shoreface being starved of siliciclastic sediment from up-dip regions, with only minimal input from local sediment supply and longshore drift (Fig 2.12).

Low magnitude fluctuations in relative sea level also had had the potential to cause similar exposure or denudation, although the geographic dislocation of each event was less than during the high magnitude events. Within passive margins low magnitude fluctuations in relative sea level have the potential to produce type 2 sequence boundaries or parasequences. Conversely, within cratonic regions relative sea level fluctuations of a similar magnitude may cause type 1 sequence boundaries and thin sequences due to the low angle slope of the basin floor. The result of composite sea level fluctuations in slowly subsiding cratonic basins is to produce a stratigraphic succession dominated by unconformity surfaces that represent a high percentage of the overall stratigraphy and facies deposited during the HST, FSST and LST (Fig 2.13).

The driving mechanisms for subsidence within cratonic basins are difficult to establish and compare with passive margins. During the Devonian there is no evidence that the North African part of Gondwana, and therefore the Murzuq Basin region, was subject to rift driven subsidence. There is also no direct evidence for large amounts of thermal or compaction related subsidence in the study region during the Devonian. The input of isostatic loading by sediment and water on subsidence will also be greatly reduced in basins underlain by old, stable continental crust when compared to basins underlain by rheologically weaker material such as oceanic crust. As such the creation of accommodation space in this area during the Devonian is unlikely to be controlled by these factors, possibly indicating that global eustatic sea level changes may be directly responsible for any changes in accommodation space.

2.4.8. *Sequence Stratigraphy in non-marine settings.*

The Lower to Upper Devonian succession in much of southern Libya, Egypt and Algeria contains some alluvial sediments and therefore it is important to consider the application of sequence stratigraphic concepts to non-marine sequences. Previous work on the extension of sequence stratigraphy to non-marine settings has included work by Aitken and Flint (1994), Flint et al. (1995), and Hampson (1995) amongst others, and is summarised in Emery and Myers (1996). Shanley and McCabe (1993,1994) outlined the effects of changes in relative sea level on fluvial architecture and the problems encountered within non-marine environments.

Within a longitudinal profile through marine to fluvial facies, the maximum inland point where marine influence is noted is referred to as the bayline (alluvial to paralic boundary), corresponding to the equilibrium point of Posamentier and Vail (1988). This distance is extremely variable and depends upon the gradient and discharge of the fluvial system, as well as the tidal range of the basin within which drainage occurs. In up-dip area the relative influence of climate, source area uplift, basin subsidence and eustasy vary with the effects of eustasy decreasing (Shanley and McCabe 1994). The rate at which accommodation space is created, eustasy, subsidence, and to a lesser degree, compaction, can drive changes in fluvial architecture which can be compared to coeval shoreface strata.

The effects of changes in base level can be interpreted within shoreface and alluvial facies and may be correlated between the two palaeoenvironmental settings when data permit (Shanley and McCabe 1993; 1994; Milana 1999). Shanley and McCabe (1993; 1994) highlight the inherent problems applying sequence stratigraphy to fluvial strata which include: (i) abrupt lateral facies changes, (ii) poor biostratigraphic resolution, (iii) limited absolute age dating, (iv) numerous internal erosion surfaces, (v) the absence of throughgoing marker horizons, and (vi) the general lack of continual outcrop down-dip to coeval shoreface strata. These problems are also emphasised by Blum (1993) and Miall (1996), the former author calling for future work to consider carefully that "changes in sediment supply and relative changes of base level" are "dependant variables that respond to independent climatic and or tectonic controls". Variations in fluvial discharge and sediment supply occur over short time scales that may act to re-enforce or dampen trends in base level (Blum 1993). Blum (1993) also pointed out that " at some point upstream rivers become completely independent of higher order relative changes in base level, and are responding to a tectonically controlled long term average base level erosion". This point was also raised by Miall (1997) (after work by Galloway 1989a; 1989b), who identified source area tectonism as the likely control on the location, timing, and thickness of Gulf coast Quaternary clastic wedge development, rather than deposition in a highstand systems tract (HST).

2.4.9. *Basic concepts in the application of Sequence Stratigraphy to non-marine successions.*

The profile of the basin slope is an important control on the effect that base level changes will have. With a fall in relative sea level, the former highstand shelf slope becomes the regional slope of the early lowstand coastal plain. This new regional slope will affect the style and distribution of fluvial deposits that are subsequently formed. An increase in the angle of the slope over which the channel flows will result in an increase of stream power. If this increase in stream power is sufficient there will be an increased potential for modification of the slope by erosion. Slope profiles can also influence the type of channel or channel pattern that develop on the slope (Wood et al. 1993; Emery and Myers 1996). Low angle slopes are characterised by straight or sinuous channels while steeper slopes often contain straighter channel forms. During a fall in base level across a steep slope, rapid vertical incision occurs as the fluvial systems adjust to the new lowstand base level. Deep vertical incision will form steep sided channels that inhibit lateral erosion and meandering, these incised valleys are deep with low width to depth ratios. Conversely, during a fall in base level across a low angle slope, less rapid vertical incision occurs as the shelf angle is less out of equilibrium with the coastal plain slopes (Plint 1996). Shallow vertical incision forms shallow incised valleys with high width to depth ratios, these formed as the increased stream energy is expended in lateral erosion and widening of valleys (Plint 1996). When a base level fall occurs and there is no change in the angle of slope beneath the new lowstand deposits there is an extension of the fluvial profile (Emery and Myers 1996). This extension of the fluvial profile may result in the fluvial systems altering their channel pattern and cause incision of the fine grained shelf sediments (Quirk 1996), depending upon the hydrodynamic properties of the new fluvial catchment area (changing distributary input, sediment supply and the physical properties of the new substrate). When a base level fall occurs where the slope of the old shelf is less than the coastal plain accommodation space is generated and fluvial aggradation will occur until a new equilibrium profile is developed (Plint 1996).

An equilibrium or graded profile is the smooth, concave-upward profile of a graded river that develops as a function of discharge and sediment load (Wood et al. 1993; Quirk 1996). If base level rises or falls, or the position of the river mouth is shifted horizontally then a river may aggrade or degrade its profile as a means of restoring the graded profile so that it can transport its load of sediment with a given discharge (Plint 1996). The magnitude of base level change has a major effect on how the river will respond. If there is a small lowering of base level then the river can respond by increasing its sinuosity, changing channel pattern or by increasing bed roughness. Therefore the increase in gradient is accommodated without any incision. High magnitude or rapid base level fall will result in channel incision as the flows are concentrated within the channel margins. When base level rises aggradation occurs although this occurs at a low rate because the entire valley has to be regraded, not just the channel (Plint 1996).

With this methodology in mind the data from the Murzuq Basin and other areas can be interpreted to determine their evolution during the Devonian.

Chapter 3. Outcrop log correlation and interpretation for the Lower Devonian succession of the Murzuq Basin, southern Libya.

This chapter contains data and interpretations relating to Lower Devonian sedimentary rocks (Fig. 1.4) that were studied on the SW margin of the Murzuq Basin (Fig. 1.3). Previous work in the area by Klitzsch (1969), Galecic (1984), Jakolovljevic (1984), Protic (1984), Radulovic (1984a) and Radulovic (1984b) identified two formations in the Lower Devonian succession, namely the Tadrart and the Ouan Kasa formations. This study has identified seven facies associations in the Lower Devonian succession which are discussed below. The subdivision of the facies associations was made on the basis of systematic variations in lithology, sedimentary structures, ichnofabric, and the floral and faunal content. A graphic log correlation panel for the Lower Devonian succession can be seen in figure 3.1 with the distribution of the logged sections in outcrop displayed in figure 3.2. A list of detailed study areas, logged sections and their locality numbers is given in appendix 3.1.

3.1. The Tadrart Formation (Gedinian to Emsian) and Ouan Kasa Formation (Emsian). General stratigraphic information.

The Tadrart and Ouan Kasa formations crop out on the SW margin of the Murzuq Basin and generally dip at a low angle (1-4°) towards the east. The outcrop pattern of the Lower Devonian succession is broadly oriented N-S with the low angle of dip providing 50 kilometres of E-W directed exposure in the southern part of the field study area (Fig. 3.2). The Lower Devonian outcrop pattern tapers northward before pinching out completely in the area around 25°48'.00N, 10°30'.00E (Fig. 3.2).

The Tadrart Formation, which comprises facies associations (1) to (6), is dominantly siliciclastic, mainly comprising very fine to very coarse-grained, locally granular, quartz sandstones. Previous work on the Lower Devonian succession in the Murzuq Basin by Beuf et al. (1971), Galecic (1984), Jakolovljevic (1984), Protic (1984), Radulovic (1984a), Radulovic (1984b), Clarke-Lowes (1985), and Selley (1997a) also noted that the sandstones were predominantly texturally and physically mature quartzites or orthoquartzites, usually comprising 95-100 % quartz. Rare examples of micaceous siltstones, mudstones and claystones were observed in the Tadrart Formation; these lithofacies were also noted by previous workers (Beuf et al. 1971; Galecic 1984; Jakolovljevic 1984; Protic 1984; Radulovic 1984a,b; Clarke-Lowes 1985; and Selley 1997a). These authors failed to accurately constrain the stratigraphic interval within which deposition of the Tadrart Formation occurred, identifying the Tadrart Formation as broadly Early Devonian using spore data. The lack of biostratigraphic data during the present study means that the Givetian to early Emsian age assigned to the Tadrart Formation by Selley (1997b) is used (Fig. 3.2).

The Tadrart Formation is generally overlain by the Ouan Kasa Formation throughout much of the study area (Fig. 3.2) although in the NW of the Murzuq Basin, the Ouan Kasa Formation unconformably overlies sedimentary rocks of the Akakus Formation (Bellini and Massa 1980; Protic 1984; Selley 1997a). This juxtaposition of the Ouan Kasa and Akakus formations is disputed by Massa (1998, pers comm.) who considers that the Ouan Kasa Formation is missing from the central and northwestern part of the SW margin. The Ouan Kasa Formation (Facies association (7)) comprises claystones, siltstones, sandstones, and ferruginous oolites, and is Emsian to lower Eifelian in age (Galecic 1984; Jakolovljevic 1984; Protic 1984).

3.1.1. *Characteristics of the lower contact of the Lower Devonian succession.*

In the southern part of the field study area (Fig. 3.2) the Tadrart Formation unconformably overlies sedimentary rocks of the Akakus Formation which are Wenlockian to Ludlovian in age (Fig. 1.4 and section 1.2.3). The unconformity surface can be clearly identified due to compositional differences between the two formations (Fig. 3.3) and the surface has up to 2 metres of relief (fig. 3.4). Field observations in the southern part of the field study region in this study, and undertaken by previous authors (Galecic 1984; Radulovic 1984a; Clarke-Lowes 1985; Selley 1997a), did not identify substantial tilting of the underlying Silurian succession prior to deposition of the Lower Devonian sedimentary rocks in this region (Fig. 3.2).

In the northern part of the field study area there is a clear angular unconformity between the Tadrart and Akakus formations (Fig. 3.5). In this northern region bedding planes in the uppermost Akakus Formation were measured dipping between 10°-20° towards the east (Fig. 3.5) while the immediately overlying sedimentary rocks of the Tadrart Formation dip between 1° and 4° towards the east (Fig. 3.6). The origin of this unconformity is discussed in section 1.2.3.

3.2. Facies Associations 1 and 2. General description.

Facies Association 1 (FA1) comprises massively bedded, laterally extensive single storey sheet sandstones that were commonly difficult to study in detail as the units form high, inaccessible cliff sections (Fig. 3.7). The sandstones in FA1 are medium to very coarse-grained, moderately sorted and are planar tabular and trough cross-bedded. The planar tabular cross-bedded units are up to 15 metres wide and up to 1.2 metres thick (Fig. 3.8) while the trough cross-bedded units are up to 15-20 metres wide and up to 2 metres thick (Fig. 3.9). Rare *Thalassinoides* burrow forms were also observed at locality 9. The cross-bedded sandstones in FA1 occur within erosive channels and commonly contain coarse-grained, clast-rich lags (Fig. 3.10). Soft sediment deformation is rarely present, comprising laterally discontinuous horizons that contain recumbent foresets (Fig. 3.11) and fluid escape structures (Fig. 3.12).

At locality 7 units of FA1 immediately overlying the Early Devonian/ Silurian unconformity comprise moderately to poorly sorted, very coarse-grained to granular microconglomeratic iron-stained sandstones. The sandstones at this locality also contain iron concretions, and iron-stained foresets and joint surfaces. At locality 5 FA1 contains low angle truncation surfaces that can be traced many tens of metres laterally, while at localities 7, 8, and 9 minor erosional scour surfaces are also commonly observed in the sandstones. The truncation surfaces are sub-horizontal and appear to be parallel with the regional dip of bedding planes.

Facies Association 2 (FA2) comprise thinly to medium bedded, laterally discontinuous, multi-storey channel sandstones with rare siltstones. The sandstones are fine to very coarse-grained, poorly to moderately sorted and contain minor scour surfaces, re-activation surfaces, and rare *Planolites* burrow forms. The sandstones contain planar tabular and trough cross-bedding similar to that observed in FA1 (figures 3.8 and 3.9), although the bedforms are generally smaller and laterally discontinuous. As well as these sedimentary structures at locality 5, FA2 also contains a 5 -10 centimetre thick package of low angle planar laminated sandstones. The cross-bedded sandstones commonly erosionally truncate each other, and at locality 3 a succession of planar tabular cross-bedded sandstones is incised by trough cross-bedded sandstones (Fig. 3.13). Soft sediment deformation is observed in laterally continuous horizons that contain recumbent foresets (Fig. 3.11) which are overturned towards the west at localities 1 and 5, fluid escape structures, and massive "fluidised" sandstones (Fig. 3.14). At a number of localities the sandstones contain low angle scour/truncation surfaces which can be traced many tens of metres laterally. These truncation surfaces are repeated vertically to produce a succession of cross-bedded sandstones cut every 50-100 centimetres by truncation surfaces. Bed thickness and grain-size commonly decrease away from these truncation surfaces.

The siltstones in FA2 are micaceous, pink-cream at outcrop and occur in laterally discontinuous lenses which are incised by the sandstones. The siltstones also contain rare *Tigillites* burrow forms.

3.2.1. *Palaeocurrent data from Facies Associations 1 and 2.*

Palaeocurrent data were collected from the logged sections and also from detailed study areas of FA1 and FA2 (Fig. 3.1). The palaeocurrent data from FA1 and FA2 is from planar tabular and trough cross-bedding although data from subordinate low angle planar laminated sandstones and low angle truncation surfaces were also measured in FA2.

The planar tabular data from FA1 (n=16) has a wide distribution but has a bi-modal W and NNW distribution (Fig. 3.1), while similar data from FA2 (n=83) has a wide distribution but also has a bi-modal NW and SW distribution. The data set from trough cross-bedding in FA1 is too small to allow confident interpretation (n=4). However, trough-cross bedding data from FA2 (n=40) have a wide distribution with a pronounced mode towards the NW with a vector mean of 303° (Fig. 3.1). FA2 also contains low angle planar laminated sandstones. However, the data set for these sedimentary structures is too small for confident interpretation. The numerous truncation surfaces in FA2 (n=15) have several different orientations (Fig. 3.1).

3.2.2. *Palaeoenvironmental interpretation of Facies Associations 1 and 2.*

Facies associations 1 and 2 are considered to have been deposited in a braidplain deposystem as outlined in figure 3.15. These interpretations are consistent with the interpretations of Klitzsch (1969), Beuf et al. (1971), Galecic (1984), Jakolovljevic (1984), Protic (1984), Radulovic (1984a), Radulovic (1984b), Clarke-Lowes (1985), and Selley (1997a). FA1 and FA2 have broadly similar lithological compositions which indicate a similar depositional environment for each facies association. The subdivision of FA1 and FA2 is made because of the distinct and systematic variations in bedform size and channel geometry between the genetic units. Deposits of FA1 are interpreted to have been deposited in large, relatively deep braid channels that contained large bedforms. Sedimentary rocks comprising FA2 were deposited in small(er), shallow(er) channel systems. The channel fill successions contain evidence of frequent erosional reworking of previously deposited bedforms.

Braided rivers are characterised by wide, shallow, low sinuosity channels (Cant 1982; Selley 1985) containing multiple thalwegs (Miall 1996) which form as a result of differential flow in the channel systems. The differential flow regime leads to the formation of composite braid bars separated by inter-bar channels (Cant 1982; Miall 1996). The large planar tabular cross bed cosets observed in FA1 and FA2 are interpreted as deposits formed by the migration of composite braid bars. According to Cant (1982), Kvale and Vondra (1993), and Miall (1996) these composite braid bars are interpreted to form as a result of multiple flood events which can form wide zones which are reworked during periods of low stage flow and/or subaerial exposure

(Collinson 1996). The trough cross-bedded sand cosets observed in FA1 and FA2 are interpreted to be the result of the migration of dunes or mega-ripples which migrate in inter-bar channels (Cant 1982; Lopez-Gomez and Arche 1993; Olsen and Larsen 1993; and Miall 1996). As outlined by Lopez-Gomez and Arche (1993), Browne and Plint (1994) and Miall (1996) these inter-bar channels may be the deepest parts of the channel network or areas of more persistent channel flow as suggested by Collinson (1996). Trough cross-bedded sandstones are also observed in modern ephemeral braided streams according to Williams (1971), where they form during rapid channel incision and fill events (Lawrence and Williams 1987). The geometry, frequency and distribution of the planar tabular and trough cross-bedding in FA1 and FA2 is interpreted to be the product of the interaction between flood events. Such flood events are interpreted by Cant (1982) to form complex bar morphologies. It is difficult to subdivide further the various bedforms observed in FA1 and FA2 due to the limited data on their three dimensional geometry and relationships to the channel margins. The planar tabular and trough cross-bedded sandstones in FA1 and FA2 will have formed at slightly different times as these lithofacies represent dunes formed under different hydraulic conditions (Miall 1996).

The recumbent foresets observed in FA1 and FA2 are interpreted to have formed as a result of the liquefaction and subsequent deformation of the cross-bedded sand by simple shear. The lateral continuity of these deformed intervals strongly suggests an external mechanism to cause the liquefaction such as an earthquake. This deformation process is common in fluvial deposits with the deformation occurring soon after deposition in unconsolidated sediment (Collinson and Thompson 1989). Soft sediment deformation structures of a similar type were also observed by Beuf et al. (1971), Galecic (1984), Jakolovljevic (1984), Protic (1984), Radulovic (1984a), Radulovic (1984b) and Clarke-Lowes (1985) in the Lower Devonian succession in this region.

The absence of abundant low angle planar laminated sandstones in FA1 and FA2 suggests that these flow conditions were not unconfined flows (Lawrence and Williams 1987). The numerous low angle truncation surfaces observed in FA2 may be scour surfaces which formed during repeated periods of high stage flow as described by Miall (1996). Alternately, the truncation surfaces formed during the lateral or down-dip accretion of bars or macroforms. The truncation surfaces in FA2 could be traced laterally across many tens of metres, but data on the geometry, orientation and genesis of these surfaces is inconclusive. Miall (1996) has identified lateral accretion surfaces within braidplain deposits. Surfaces of this kind were originally interpreted by Kraus and Middleton (1987) to be characteristic of meandering streams. However, braidplain deposits can comprise cosets of cross strata with no discernable lateral or down-dip accretion (Collinson 1996). Friend (1983) identified two dominant geometries of sand bodies within fluvial deposits; lenticular bedded (multistorey) sheet sandstones dominated by longitudinal progradation; tabular bodies dominated by lateral accretion. Deposits of FA1 and FA2 may comprise both sand body types but limited data on the nature of the surfaces within and bounding the sand bodies prevents subdivision using these criteria. Similarly this lack of detailed geometrical data prevents the interpretation of the fluvial sand bodies in terms of their relationship with the channel they lie within, that is, whether they are attached to the channel margin, or related in scale to the dimensions of the channel, a framework outlined by Collinson and Thompson (1989). The grain-size separation commonly observed between foresets in FA1 and FA2 is thought to be the result of small scale bedforms migrating up the stoss side of bedforms and then avalanching down the slipface of the larger 2D and 3D dunes, a process in fluvial deposits which was described by Hunter (1985).

According to Cant (1982), Selley (1985) and Allen and Allen (1990) a lack of fine-grained sediments within alluvial successions is characteristic of braidplain deposits. However, fine-grained sediment can occur within sandy braidplains as abandoned channel fill sequences (Selley 1985). The clast-rich lags at the base of many of the channels in FA1 and FA2 (figures 3.1 and 3.10) are interpreted to be the result of the erosional reworking of the limited amounts of fine-grained sediment that was deposited in overbank areas and abandoned channels (Fig 3.15). The Lower Devonian braidplain on the SW flank of the Murzuq Basin may be analogous to the

Ivishak Formation of Alaska (Triassic) where over 200 metres of braided fluvial sandstones are observed, within which shale lenses are uncommon (Cant 1982).

The presence of a rare, monotype *Thallasinoides* ichnofabric in FA1 and rare examples of a monotype *Planolites* or *Tigillites* ichnofabric in FA2 may indicate that paralic palaeoenvironmental conditions were rare. However, limited data from lateral sections means the significance and regional distribution of this ichnofabric is poorly understood.

The sedimentary structures and alluvial architecture of FA1 and FA2 are similar to the Platte-type braided river of Miall (1977), and the shallow, perennial, sand-bed braided river of Miall (1996). Both of Miall's braidplain models are characterised by large planar, tabular cross-bed sets and subordinate trough cross-bed sets, although the latter bedform type are more common than in either of the two models, possibly suggesting highly mobile channel margins. The clast-rich channel lags in FA1 (figure 3.10) are interpreted to correspond with the reworked fine-grained floodplain sediments in the Platte type braided river.

3.2.3. *Reconstructing palaeochannel morphologies during the deposition of Facies Associations 1 and 2.*

Palaeocurrent data from planar tabular cross-bedding in FA1 and FA2 generally have a bi-modal distribution and are bisected by the vector mean from the trough cross-beds (Fig. 3.1). The distribution of palaeocurrent data from planar tabular cross-bedding in FA1 and FA2 is interpreted to be the result of the migration of straight crested braid bars which migrated transverse to the primary flow direction (figures 3.15 and 3.15a.). Straight crested bars can migrate in any direction between normal and perpendicular to the main flow (Lopez-Gomez and Arch 1993; Miall 1996). In FA1 and FA2 planar tabular cross-bedding indicate the migration of straight crested braid bars towards the W-SW and N (Fig. 3.1). During the deposition of FA1 and FA2 the inter-bar channels contained fields of 3d dunes which migrated normal to the primary flow direction resulting in the deposition of trough cross-bedded sandstones (figures 3.15 and 3.15b). In FA2 trough cross-bedding indicates that 3d dunes in the braid channels generally migrated towards the NW. The overall distribution of palaeocurrent data suggests that bedforms were migrating in low sinuosity channels which were generally oriented towards the NW (Fig. 3.15). The direction of braided fluvial drainage identified in this study corresponds to that inferred by Beuf et al. (1971), Jakolovljevic (1984), Clarke-Lowes (1985), and Selley (1997a).

Braided channels commonly fill by vertical accretion during waning flow and are capped by small 2D and 3D dunes and ripples modified during low stage conditions and subaerial exposure (Miall 1996). However, the complete waning flow fill of a channel is rarely seen due to subsequent reworking (Cant 1982). The absence of waning fill successions in FA1 and FA2 is interpreted to be the result of frequent erosional reworking of the channel fill units to form complex coalesced bar forms.

The channel systems of braided rivers are laterally unstable because of the lack of cohesive floodplain sediments and generally high discharge peaks that result in the lateral migration of channels. This lateral migration can result in the preservation of a laterally continuous sand body that may contain a number of diachronous and coeval facies associations which lie in clearly or poorly defined channels. According to Coleman (1969) within the modern Brahmaputra river lateral migration rates of several thousand metres in a single flood are not uncommon. Campbell (1976) identified palaeo-braided fluvial channels in the Morrison Formation of New Mexico (Jurassic) up to 11 kilometres across which contain several coalesced, smaller channels. The bounding surfaces of these large channels can be very low angle, sloping at a few degrees or less (Miall 1996). Channel scour surfaces can place identical sandstones in contact with each other (Miall 1996), and as a result these surface can be difficult to identify in the field. Laterally migrating, mobile channels generally produce successions dominated by cross-bedded units, rather than the massive sandstones dominated by upper flow regime sedimentary structures which result from unconfined sheet floods (Allen and Allen 1990). The laterally continuous, thinly to thickly bedded sheet sandstones of FA1 and FA2 are interpreted to be the product of laterally migrating mobile channel belts rather than by unchannelised sheet floods.

Miall (1996) pointed out that the identification of channel margins in laterally migrating sheet sandstones may be difficult or impossible to identify. The dimensions of the channels within which the bedforms of FA1 and FA2 were migrating were not determined during the course of this study, possibly due to the above factors or partly due to the limited availability of sections oriented perpendicular to the flow direction. However, the increased frequency of erosion surfaces and smaller bedforms in FA2, relative to FA1, are thought to indicate a change in the hydrodynamic properties of the alluvial systems. The increased reworking and smaller bedforms may indicate a greater degree of topographic differentiation within the alluvial system, with more numerous, smaller channels which are relatively narrower with steeper sides.

3.2.4 *Interpretation of palaeoclimatic indicators and the basinal setting during the deposition of Facies Associations 1 and 2.*

The regional palaeoclimate in the Murzuq Basin during the Early Devonian (section 1.4) may have promoted the formation of braided rivers. According to Schumm (1969) during the Early Devonian terrestrial plants were in an early stage of their evolution and there would have been limited vegetation cover. As noted by Miall (1996) the braiding of river channels is normally associated with rivers with emphemeral or highly irregular fluvial discharge and a high bedload content. The sparse vegetation cover interpreted in the region during the Early Devonian probably had little effect on impairing rapid run-off after rainfall, resulting in highly irregular discharge rates. It is also considered that the amount of bedload transported within fluvial systems was much greater due to the lack of vegetation and variable discharge, factors which, according to Schumm (1969), Allen and Allen (1990), and Miall (1996) promote braiding in fluvial systems. Miall (1996) suggested that the Early Devonian land surface away from the active braid channels probably resembled modern arid areas, even where rainfall was high. The irregular and high volume run-off, and generally sandy surface cover, probably meant that inter-fluve palaeosols were well drained and formed when the rate of run-off decreased due to seasonal or longer time scale climatic cycles. However, the preservation potential of palaeosols in successions dominated by such braidplains may be low due to the frequent lateral incision and avulsion of active channels successions (Miall 1996) (Fig. 3.17). As terrestrial plants evolved, colonised and spread during the Devonian the preservation potential of palaeosols and fine-grained sediments increased as the sediment was stabilised by the vegetation cover (Miall 1996). Longer time scale palaeoclimate cycles driven by eustasy and tectonism, including rainfall patterns, may also have promoted the formation and preservation of extensive palaeosols. Although no evidence was found in this study, plant fragments and spores do occur in the Tadrart Formation on the western margin of the Murzuq Basin, indicating the colonisation of continental areas on the northern margin of Gondwana by primitive plants (Klitzsch 1969; Beuf et al. 1971; Galecic 1984; Jakolovljevic 1984; Protic 1984; Radulovic 1984a; Radulovic 1984b; Clarke-Lowes 1985; Selley 1997a). This organic material may be derived from vegetation that colonised some bar forms, creating channels that were temporarily stable with a reduced tendency towards braiding (Schumm 1969; Selley 1985; and Miall 1996).

As noted by Cant (1982) it is not uncommon for aeolian deposits to be closely associated with fluvial systems which can provide an abundant source of sand. He also pointed out that there may be some component of aeolian re-working of fluvial deposits during low or falling river stages. The identification and differentiation of deposits that formed by the migration of 2D and 3D dunes in fluvial or aeolian environments can be problematic according to Cant (1982), and as such aeolian deposits cannot be completely discounted in FA1 and FA2. Similarly, it is commonly difficult to define the limit between the fluvial and deltaic systems when braid deltas are present (Coleman and Prior 1982). It may be that some of the sedimentary rocks in FA1 and FA2 were deposited on the upper delta plain rather than in an up-dip fluvial setting. However, this palaeoenvironmental distinction cannot be made due to the poorly understood regional three dimensional inter-relationship between the two facies associations and the position of the palaeoshoreline and delta down-dip.

According to Coleman and Prior (1982) the distinction between braided fluvial and braid delta deposits can be problematic as the two environmental settings contain very similar depositional processes and resultant deposits.

A number of outcrop and regional variations in bed thickness and bedform inter-relationships are observed in FA1 and FA2. These variations in alluvial architecture may be the result of allocyclic or autocyclic controls, which are addressed in section 3.9.1.

3.3. Facies Association 3. General description.

Facies association 3 (FA3) comprises laterally discontinuous thinly to massively bedded fine siltstones and mudstones with thin sandstone interbeds. The fine siltstones and mudstones are cream or red and contain sub-angular to sub-rounded clasts of fine to very coarse-grained sandstone. The thinly bedded sandstones are laterally discontinuous, fine to very coarse-grained, poorly to well sorted and contain mudstone flakes, rounded clasts, and plastically deformed clasts of siltstone or mudstone (Fig. 3.16). The sandstones also contain an ichnofabric of low to medium abundance and low diversity, comprising *Planolites* and an unidentified, large horizontal trace (Fig. 3.17). The sandstones are planar tabular and trough cross-bedded, the upper beds of which contain symmetrical ripples (Fig. 3.1). The dominantly fine-grained sedimentary rocks of FA3 (Fig. 3.18) are very uncommon in lateral and vertical profiles through the Lower Devonian succession and are wholly encased in units of FA1 and FA2.

At locality 9, sedimentary rocks comprising FA3 form a 4 metre thick lenticular body, pinching out laterally across 20 metres of exposure. The basal contact of FA3 at this locality is a concave-upward scour surface that incises into the underlying sandstones. The fine siltstones and muds at locality 9 generally weather to a cream colour although where they are in contact with the under-and over-lying sandstones they are red (Fig. 3.18). There are also large red nodules within the cream mudstone (figures 3.1, 3.15c and 3.18).

3.3.1. *Palaeocurrent data from Facies Association 3.*

Palaeocurrent data from FA3 were collected from the logged section at locality 9 (Fig. 3.1) and comprise measurements from thinly bedded planar tabular and trough cross-bedding and symmetrical ripples in the sandstones. The dataset from the planar tabular and trough cross-bedded sandstones is too small for confident interpretation. The symmetrical ripples have straight crests oriented N-S and NE-SW, indicating oscillatory flow conditions perpendicular to the ripple crests.

3.3.2. *Palaeoenvironmental interpretation of Facies Association 3.*

Facies association 3 is interpreted to comprise abandoned channel fill and laterally discontinuous flood plain deposits (Fig. 3.15c), concurring with the interpretations of Galecic (1984), Jakolovljevic (1984), Protic (1984) and Radulovic (1984a,b). The red and cream colouration of the mudstones at locality 9 is thought to be the result of the variable oxidation state and/or organic content of the sedimentary rocks, altered by a variety of pedogenic processes (Miall 1996). At locality 9 sedimentary rocks comprising FA3 occur within a concave up scour interpreted as a braid channel (Fig. 3.15c). This channel was scoured during high flow conditions and subsequently filled by the terminal fill of an abandoned braid channel. This abandoned channel may be lateral to the active channel complex, or may have been filled by the process of reverse eddy transport (cf. Selley 1985). The ichnofabric in FA3 indicates more favourable conditions for colonisation by burrowing organisms in sub-environments of FA3 than during the deposition of FA1 and FA2. These more favourable palaeoenvironmental conditions may be been the result of a transition from a braidplain to paralic setting (section 3.9.1). Symmetrical ripples can form as a result of oscillatory flow conditions within a variety of marine to continental sub-environments (Selley 1985), and therefore the symmetrical ripples in the sandstones in FA3 are not diagnostic of any specific palaeoenvironment. The dominantly fine-grained succession of FA3 has a low preservation potential as it occurs in conjunction with high energy braidplain deposits of FA1 and FA2 detailed above. The preservation of the channel fill complex comprising FA3 at locality 9 may therefore be an isolated example of what was a common lithofacies.

3.4. Facies Association 4. General description.

Facies Association 4 (FA4) comprises laterally continuous, erosively-based thinly to medium bedded mudstones, siltstones and sandstones (Fig. 3.1). The mudstones and siltstones are red and contain sub-angular and sub-rounded clasts of fine to very coarse-grained sandstone. The thickness of the mudstone and siltstone beds commonly varies laterally, having been incised by the overlying sandstone beds (Fig. 3.19). The sandstones are fine to coarse-rarely very coarse-grained, and are moderately to well sorted. The sandstones are commonly internally structureless with rare symmetrical ripples, planar tabular and trough cross-bedding, low angle planar laminae, and soft sediment deformation (fluid escape structures and convolute bedding). The upper surfaces of many of the sand beds are irregular and iron-stained. The sandstones also contain an ichnofabric of low to medium diversity comprising *Planolites* and *Thalassinoides.*

3.4.1. *Palaeocurrent data from Facies Association 4.*

Palaeocurrent data from FA4 were collected from the logged sections and detailed study areas (Fig. 3.1). The palaeocurrent data comprises measurements from occasional thinly bedded planar tabular and trough cross-bedded and low angle planar laminated sandstones with symmetrical ripples. The data set from the low angle planar laminated, planar tabular and trough cross-bedded sandstones is too small for confident interpretation. The symmetrical ripples have straight crests oriented N-S, NW-SE and NE-SW indicating oscillatory flow conditions perpendicular to the ripple crests.

3.4.2. *Palaeoenvironmental interpretation of Facies Association 4.*

Facies Association 4 is interpreted to have been deposited in coastal plain and flood plain environments, laterally or down-dip of FA1 and FA2 (Fig. 3.20). The palaeoenvironmental interpretations for FA4 correspond to those proposed by Galecic (1984), Jakolovljevic (1984), Protic (1984), Radulovic (1984a) and Radulovic (1984b). The increased amounts of fine-grained material indicates a decrease in mean depositional energy when compared to FA1 and FA2, which may be due to a change in the fluvial channel style from braided to anastomosing or deposition in areas away from the main channels. The thinly bedded sandstones in FA4 are structureless or contain low angle planar laminae, fluid escape structures, and convolute laminae, sedimentary structures which indicate a high rate of deposition (Selley 1985). The iron-stained upper surfaces to many of the sandstones in FA4 are interpreted to be primary features that resulted from frequent periods of subaerial exposure during times of non-deposition. These traits imply that many of the sandstones in FA4 were deposited in overbank areas lateral to the main fluvial channel network (Fig. 3.19). The thinly bedded sheet sandstones in FA4 with symmetrical ripples may represent sediment reworking by wave action in a paralic setting.

The ichnofabric in FA4 indicates more favourable conditions for colonisation by burrowing organisms in FA4 compared with FA1 and FA2. These conditions, recorded in inter-channel areas, may represent a change from the alluvial conditions of FA1 and FA2 to a paralic environment. However, the in-channel paralic facies deposited laterally or down-dip of units of FA4 were not confidently located during this study.

3.5. Facies Association 5. General description.

Facies Association 5 (FA5) comprises laterally continuous thinly to medium bedded sandstones with rare siltstones. The sandstones are very fine to very coarse-grained, poorly to moderately sorted, with the very fine to fine-grained sandstones being rarely micaceous. The sandstones contain weathered-out rounded clasts, siltstone/mudstone flakes and an inchnofabric of high abundance comprising *Planolites*, *Tigillites*, *Skolithos*, *Cruziana*, *Arthophycus*, *Diplocraterion* and possibly *Gordia* and *Ophiomorpha*. At localities 2, 5 and 6, lateral variations in the diversity and abundance of the ichnofabric are common, with beds dominated by the *Skolithos*

(Fig. 3.21) or *Cruziana* ichnofabric (Fig. 3.22). At locality 4, trough cross-bedded sandstones are erosionally truncated by bioturbated sandstones. At locality 6, *Skolithos* burrows penetrate a 40 centimetre thick sandstone bed into the underlying siltstone, this succession can be traced over 150 metres of lateral exposure and occurs within a low angle convex-up scour surface.

The sandstones of FA5 are usually erosively-based, bound by laterally continuous planar and convex-up scour surfaces. At locality 5 a laterally continuous scour surface is covered by weathered-out clasts, *Planolites* burrows, and rare symmetrical ripples. The sandstones also contain laterally discontinuous scour surfaces, re-activation surfaces, cross-cutting planar tabular and trough cross-bedded cosets, and low angle planar laminated sandstones. At locality 2 the cross-bed cosets are up to 1.5 metres high and can be traced laterally over tens of metres. The thinly bedded sandstones contain asymmetrical ripples, ripple cross-laminae, symmetrical ripples and contain rare fluid escape structures. At locality 5 symmetrical ripples are observed. At localities 2 and 6 the upper surfaces of some of the sandstones are iron-stained.

3.5.1. *Palaeocurrent data from Facies Association 5.*

Palaeocurrent data from FA5 were collected from the logged sections and detailed study areas (Fig. 3.1). The palaeocurrent dataset from FA5 was obtained from planar tabular and trough cross-bedded sandstones with subordinate data from low angle planar laminatae, ripple cross-laminae and symmetrical ripples. The planar tabular cross-bedding data from FA5 (n=51) has a wide distribution but has a poorly defined bimodal W and NNW distribution (Fig. 3.1). The trough cross-bedding data from FA5 (n=63) has a wide distribution but the pronounced mode gives a vector mean of 309°. The palaeocurrent dataset from the low angle planar laminated sandstones is too small for confident interpretation. The ripple cross-laminated sandstones were formed by the migration of asymmetrical ripples towards the south (towards 180° and 192°) and N-NW (towards 342°, 344° and 000°). The symmetrical ripple crests are oriented N-S, NE-SW and NW-SE and were formed by oscillatory currents oriented perpendicular to the ripple crests.

3.5.2. *Palaeoenvironmental interpretation of Facies Association 5.*

Facies Association 5 is interpreted to have been deposited within marine influenced braidplain and braid delta palaeoenvironments (Fig. 3.23) consistent with the interpretations of Galecic (1984), Jakolovljevic (1984) and Protic (1984). The sand dominated succession and clast-lined scour surfaces indicate that the mean depositional energy was high and that erosional re-working of fine-grained material occurred. The diverse and abundant ichnofabric indicates favourable conditions for colonisation within the sediment and above the sediment/water interface. These conditions are generally associated with an increase in the marine influence acting within and upon the sediments, in marginal to shallow marine conditions. The incision of non-bioturbated sandstones by bioturbated sandstones shown in figure 3.24 is interpreted to indicate the erosional re-working of fluvially generated bedforms by tidal currents as alluvial discharge decreased or the main fluvial channels migrated laterally (Fig. 3.23). The laterally discontinuous *Skolithos* bioturbated sandstones may also represent marine influenced channel deposits that incised sandy alluvial, delta plain or fluvial deposits (Fig. 3.23). The laterally discontinuous iron-stained surfaces observed on the upper surfaces of some of the sandstones are interpreted to have formed during short periods of subaerial exposure. These periods of subaerial exposure occurred as the rate, and location of active deposition in fluvial, deltaic, and marginal marine environments varied.

Palaeocurrent data from planar tabular cross-bedding in FA5 is bimodal towards the WSW and NNW indicating the migration of straight crested bars in these directions. These straight crested bars are interpreted to be migrating in a downstream direction (Fig. 3.25) and may represent composite braid bars separated by inter-bar channels (Cant 1982) (Fig. 3.23). The distribution of planar tabular cross-bedding palaeocurrent data from FA5 is broadly comparable to palaeocurrent data from similar bedforms in FA1 and FA2. The trough cross-bedded sandstones in FA5 are interpreted to form in inter-bar channels during periods of increased discharge in a process similar to that described by Miall (1996) in fluvial deposits. The trough cross-bedded sandstones in FA5 formed by the migration of sinuous crested bars or linguoid bedforms, primarily towards the NW (Fig. 3.26). The distribution of trough cross-bedding palaeocurrent data from FA5 is similar to palaeocurrent data from the same bedforms in FA2, indicating channel networks with a similar orientation within which bedforms migrated in comparable directions.

According to Miall (1996) the absence of thick cosets of cross-bedded sandstones and complex stacked sand bodies within alluvial and paralic successions suggest a low degree of topographic differentiation within the channel network. Such sedimentological features are absent from the sandstones in FA5 suggesting that the channels were wide and shallow with poorly defined inter-channel bars.

The current ripples overlain by symmetrical ripples in FA5 are interpreted to have formed initially by uni-directional current activity and were later modified by oscillatory flow conditions. Flow conditions of this type can occur within fluvial and deltaic channels that are subject to marine processes, often within the tidal range. The symmetrical ripples are the only sedimentary structures in FA5 that may indicate oscillatory marine currents. However, symmetrical ripples can also result from the interaction of currents derived from fluvial, aeolian and marine systems. The interaction of alluvial and marine processes such as alluvial discharge, wave action and tides, can result in separate periods when fluvial and marine currents dominate depositional processes. The inland limit of sedimentary structures influenced by marine processes can be many kilometres inland of the limit of saline intrusion (Dalrymple et al. 1992; Allen and Posamentier 1993). The angle at which the basin profile dips can strongly effect the amount of, nature of, and inland extent of the marine influence on a coastal plain with low angle dips promoting marine incursions (Wright and Coleman 1973). As outlined in section 1.4.1, during the Lower Devonian the basin profile sloped at a low angle towards the NW, with no shelf slope break. Therefore, in the Murzuq Basin during the Early Devonian the up-dip limit of marine influence may have been many kilometres up-dip of the coeval palaeoshoreline.

3.6. Facies Association 6. General description.

Facies Association 6 (FA6) comprises thinly to massively bedded sandstones with siltstones and mudstones. The sandstones are fine to upper medium-grained, occasionally microconglomeratic, poorly to well sorted, and contain an ichnofabric of low to high abundance comprising *Planolites*, *Tigillites*, *Skolithos*, *Chondrites*, and *Gordia*. At localities 4 and 7 sandstones commonly contain a monotype ichnofabric of *Skolithos*. The sandstones are commonly devoid of sedimentary structures but ripple cross-laminae, laterally discontinuous planar tabular and trough cross-bedding, low angle planar laminae, and fluid escape structures are observed. At locality 11, FA6 is dominated by purple stained fine-grained sandstones which are interbedded with well sorted medium sandstones containing low angle planar laminae and symmetrical ripples. The upper beds of many of the sandstones at locality 7 are iron-stained and contain asymmetrical or symmetrical ripples. The sandstones commonly incise into the siltstones and mudstones. At locality 6 the siltstones and mudstones occur as laterally discontinuous lenses and are interbedded with fine-grained sandstones.

3.6.1. *Palaeocurrent data from Facies Association 6.*

Palaeocurrent data from FA6 were obtained from occasional planar tabular and trough cross-bedding, ripple cross-laminae, low angle planar laminae and asymmetrical and symmetrical ripples (Fig. 3.1). The dataset from planar tabular and trough cross-bedding is limited (n=7 and n=11 respectively). However, the distribution of the data (Fig. 3.1) are broadly comparable to the distribution of data from FA1, FA2, and FA5. The dataset from low angle planar laminated sandstones is too small for confident interpretation. The ripple cross-laminated sandstones were formed by the migration of asymmetrical ripples towards the south (towards 175°, 189° and 192°) and N-NW (towards (325°, 340° and 347°). The symmetrical ripples have crests oriented N-S and NE-SW that formed as a result of oscillatory current action oriented perpendicular to the ripple crests.

3.6.2. *Palaeoenvironmental interpretation of Facies Association 6.*

Facies association 6 is interpreted to have been deposited within marine influenced braid delta, interdistributary bay, tidal flat, tidal channel, and shoreface palaeoenvironments (Fig. 3.23), corresponding to the interpretations of Galecic (1984), Jakolovljevic (1984) and Protic (1984). The variations in ichnofabric observed in FA6 are interpreted to be the result of subtle changes in the physical and chemical conditions within the sub-environments outlined above. The laterally discontinuous sandstones that contain monotype *Skolithos* ichnofabrics are interpreted as tidal channels and shoreface deposits (Fig 3.23). The bioturbated sandstones and siltstones are interpreted as tidal flat deposits, these are occasionally incised by tidal channels which contain planar tabular and trough cross-bedding, formed by the migration of straight and sinuous crested bedforms (Figures 3.15a and 3.15b respectively). The laterally discontinuous iron-stained upper surfaces observed on some of the sandstones are interpreted to have formed during short periods of subaerial exposure, resulting from variations in the rate of deposition and location of fluvial, deltaic, and shallow marine depositional systems.

Palaeocurrent data from planar tabular and trough cross-bedded sandstones in FA6 is of limited interpretative use due to the small data set available. However, the distribution of planar tabular and trough cross-bedding data indicates that bedforms were generally migrating towards the W or N in these channels, down the regional dip (Fig. 1.12). The distribution of palaeocurrent data from asymmetrical ripples indicate that during low flow speeds current flow direction was variable (Fig. 3.1). These low flow speeds, as defined in figure 5.14 of Miall (1996), can occur within channels or on larger bedforms during low or falling stage conditions, or may result from interacting wind, fluvial, deltaic, or ebb and flow tidal generated currents (Fig. 3.23). The palaeocurrent data from FA6 indicates that the effects of marine derived currents e.g. wave action, were limited, probably as a result of the basin profile (section 1.4.1), although high rates of fluvial discharge can also subdue wave action and form fluvially dominated deltas (Wright and Coleman 1973).

3.7. Facies Association 7. General description.

Facies Association 7 (FA7) comprises laterally discontinuous thinly to medium bedded claystones, ferruginous oolitic siltstones and sandstones (Fig. 3.27). At locality 11 the oolitic sandstone bars pass laterally into siltstones and mudstones before being replaced by successive oolitic sandstone bars. The siltstones contains between 5 and 20 % oolitic material while the sandstones contain between 0 and 70 % oolitic material, the latter being fine to medium-grained, sub-rounded and ferruginous. The sandstones are fine-grained and contain rounded siltstone clasts, shell debris, shell moulds, and an ichnofabric of low to medium diversity comprising *Planolites*, *Tigillites*, and *Chondrites*. The sandstones occur in erosively-based beds within which the grain-size commonly fines or coarsens-upward. The sandstones contain rare planar tabular and trough cross-bedding at locality 11, low angle planar or parallel laminae, soft sediment deformation or have no internal structure. The upper surfaces of many of the sandstones are iron-stained and irregular.

Geochemical analysis of the claystones in FA7 (Ouan Kasa Formation) by Galecic (1984), Jakovljevic (1984) and Protic (1984) stated that they mainly comprised kaolinite with subordinate illite-montmorillonite, alunite, jarosite, gypsum, goethite and quartz.

3.7.1. *Palaeocurrent data from Facies Association 7.*

Palaeocurrent data from FA7 were obtained from planar tabular and trough cross-bedding with the oolitic sandstones with subordinate data from low angle planar laminae. The dataset from the planar tabular and trough cross-bedding is of limited size (n=10 and n=12 respectively) but may be representative of the variable palaeocurrents active during the deposition of FA7 (Fig. 3.1). The dataset from the low angle planar laminated oolitic sandstones is also of limited size (n=7) but demonstrates that southeasterly directed currents were active during the deposition of FA7 (Fig. 3.1).

3.7.2. *Palaeoenvironmental interpretation of Facies Association 7.*

Facies Association 7 is interpreted to have been deposited within shallow marine and lagoonal palaeoenvironments (Fig. 3.28). The lateral variations in litholofacies are interpreted to have resulted from transitions between offshore bar, shoreface, tidal flat, and lagoonal conditions. The interpretations correspond to those of Galecic (1984), Jakolovljevic (1984) and Protic (1984) who suggested a littoral to sublittoral palaeoenvironment during the deposition of the Ouan Kasa Formation.

The ooliths in FA7 occur in variable amounts within the cross-bedded sandstones and are interpreted as offshore bar deposits that migrated in shallow waters across a shallow shelf. These palaeoenvironmental interpretations for FA7 broadly correspond to the palaeoenvironmental interpretations of Clarke-Lowes (1985) and Guerrak (1991). The ooliths occurring within the siltstones are interpreted to have been deposited in inter-bar areas and low energy areas behind the oolite bars. The fining-upward of grain-size noted in several of the sandy oolite beds is interpreted to result from the migration of sandy/oolitic dunes which are separated by fine-grained inter-dune areas (Fig. 3.28).

The thinly bedded quartz sandstones, which are interbedded with siltstones and mudstones, are interpreted to be the result of occasional storms or strong tidal activity. The mud clasts, siltstones clasts, and shell debris in the sandstones also indicate a high depositional energy for these thinly bedded sandstones. The pitted upper surfaces of many of the thinly bedded sandstones are interpreted to be the result of bioturbation during low energy conditions and/or the result of oxidation and/or subaerial exposure. Similar iron-stained surfaces in FA7 (Ouan Kasa Formation) were observed by Clarke-Lowes (1985) who interpreted them to be the result of the subaerial exposure of tidal flats. Clarke-Lowes also observed desiccation cracks in the Ouan Kasa Formation which were taken to indicate sub-aerial exposure of these units.

Palaeocurrent data from the planar tabular and trough cross-bedded oolitic sandstones is of limited use in determining bedform migration directions but indicate that straight and sinuous crested bedforms were present. The low angle planar laminated oolitic sandstones are interpreted to have formed during upper plane bed conditions, the southeasterly directed data from these sedimentary structures indicates storm activity or possibly the flow of tidal currents in this direction.

Previous studies of Devonian ferruginous oolites in the Murzuq Basin and surrounding region have proposed a variety of models to explain their formation and occurrence (Bennacef et al. 1971; Guerrak 1991; Pierobon 1991; Selley 1985; 1988; Van Houten and Karasek 1981). The general consensus is that oolites form by the bonding of aragonite crystals around a nuclei (Selley 1988) which is often bioclastic (Selley 1985). The physiochemical environment which promotes oolith formation is poorly constrained but recent ooliths occur in conjunction with blue-green algae which may play some role in aragonite precipitation (Selley 1985). Modern oolites occur in high energy environments such as sand banks and tidal deltas and it was noted by Selley (1985; 1988) that oolites generally formed where cool dilute sea water mixes with warm waters within lagoons and restricted shelves. Ferruginous oolites are interpreted to accrete in quiet conditions (Chauvel and Guerrak 1988; Guerrak 1991 and references therein) forming within iron rich muds (Guerrak 1989; 1991). The ferruginous oolites in FA7 generally occur in a quartz rich matrix (silt to medium-grained sand) and thus fall within the definition of the Ferruginous Oolite Detrital (FOD) ironstone facies of Guerrak (1991). The geographic distribution and thickness of the oolitic ironstones in FA7 (Ouan Kasa Formation) in the Murzuq Basin are of the Local Ironstone Deposition (LOID) type of Guerrak (1989; 1991). The source of iron within ferruginous oolites may be localised, as is generally noted with LOID deposits (Guerrak, 1989; 1991), but in the case of the ferruginous oolites in FA4 a remote source is postulated by Guerrak (1991). This remote source of iron may be the West African craton, the Nigerian Pan-African chain, and the Congo shield (Guerrak 1991). However, palaeocurrent data from the Lower Devonian detailed above, and palaeogeographic reconstructions for the pre-Lower Devonian (section 1.3), suggest a southeasterly clastic source. The iron in FA7 is thought by Guerrak (1989) to have been transported down rivers from weathered basement rocks. However as noted in this research, and previous work by Beuf et al. (1971), Bellini and Massa (1980), Galecic (1984), Jakolovljevic (1984), Protic (1984) and Radulovic

(1984a; 1984b), the Lower Devonian Tadrart and Ouan Kasa formations unconformably overlie sedimentary rocks of Ordovician and Silurian age. The sedimentary rocks immediately underlying the Lower Devonian succession contain numerous ferruginous-stained surfaces, nodules and iron cemented horizons, particularly in the Silurian Akakus Formation, which would have provided an ample source of iron. The palaeolatitude of the study region during the Early Devonian (section 1.3) places these oolitic ironstones in cool to temperate mid-latitudes, with the source of iron in colder latitudes much further south (Guerrak 1989).

According to Galecic (1984), Jakolovljevic (1984) and Protic (1984) geochemical analyses of clays in FA7 suggests that they were deposited within the quiet brackish waters of a closed basin, probably a lagoon within which waters were acidic with a pH from 2-6.

According to Massa (1998, pers comm.) The Ouan Kasa Formation (FA7) is absent from the northern part of the outcrop belt on the SW flank of the basin (Fig. 3.2). The possible reasons for this absence are discussed in chapter 5.

3.8. Sequential facies evolution during the Early Devonian.

The facies associations outlined above were studied along the western flank of the Murzuq Basin and form a vertical profile (Fig. 3.29) which can be used to interpret changes in fluvial architecture, base level, and palaeogeography. The vertical profile illustrated in figure 3.29 shows the thick stratigraphic succession in the central part of the study area and also the attenuated succession in the northern part of the study area. The possible origins of the thickness variations observed in the Lower Devonian succession are discussed in chapter 5.

Within the lower part of the Tadrart Formation the succession is dominated by braidplain or deltaic channel sandstones and rare inter-channel sandstones of facies associations 1 and 2, with subordinate interbeds of the overbank, flood/delta plain and abandoned channel mudstones, siltstones and sandstones of facies associations 3 and 4 (Fig. 3.29). In the upper part of the Tadrart Formation there is increasing evidence of marine influence, with a succession of marine influenced braidplain, deltaic and shallow marine sandstones with rare siltstones and mudstones from facies associations 1, 2, 5 and 6. These marine influenced siliciclastic sedimentary rocks are overlain by the fine-grained, oolitic lagoonal to shallow marine deposits of facies association 7 in the central and southern part of the field study area (Fig. 3.29), although these sedimentary rocks are absent from the NW of the basin (Fig. 3.2).

3.9. Sequence Stratigraphic interpretation for the Early Devonian.

The Lower Devonian succession can be interpreted using the sequence stratigraphic methods detailed in section 2.4, with specific reference to non-marine sequence stratigraphy. As outlined above the Lower Devonian succession was deposited in shallow marine and continental conditions. The juxtaposition of Lower Devonian facies in the Murzuq Basin can be used to interpret base level and/or relative sea level fluctuations. As stated in section 2.4 the application of sequence stratigraphy to non-marine successions, and in this case within an intracratonic basin, can be problematic. However, the resulting predictive and interpretative framework can be very useful if used in conjunction with other analytical techniques (Quirk 1996).

The facies associations outlined above, and illustrated in figure 3.29, occur within linked systems tracts, within which the internal juxtaposition of facies and variations in alluvial architecture are driven by changes in the rate of creation of accommodation space and resultant relative sea level. These changes can be used to interpret the sequence stratigraphic significance of the Lower Devonian succession and interpret changes in the rate of creation of accommodation space. The driving mechanisms responsible for these changes in accommodation space are outlined below.

3.9.1. *Sequence Stratigraphic interpretation for Facies Associations 1 and 2.*

Facies Association 1: The channel fill sandstones in FA1 occur in large, wide, single storey channels which indicate that after channel incision, a relatively large amount of accommodation space was present within and laterally to the active channel systems

(Fig. 3.30). The geometry and distribution of bedforms in the single storey sandstone bodies of facies association 1 is controlled by autocyclic factors. The lateral continuity of these sandstone bodies, and little or no erosional re-working of the channel fill complexes, indicates moderate to high amounts of accommodation space.

The channel systems in FA1 amalgamate laterally to form continuous sandstone sheets which suggests that once accommodation space within an active channel scour filled, the channel system migrated laterally rather than reworking the channel fill sandstones. The lateral migration of these fluvial channels suggests that accommodation space was present throughout the region (Fig 3.30). These increased amounts of accommodation space can occur in a number of ways in fluvial systems (autocyclic vs allocyclic), but the limited data from coeval facies downdip means that it is difficult to establish the precise cause. It is possible for fluvial aggradation to result from deltaic progradation (Emery and Myers 1996). However, the profile of the Murzuq Basin makes this mechanism unlikely (Fig. 1.12; section 1.4.1). Variations in the rate of subsidence can also cause aggradation to occur (Wehr 1993) with the preservation of thick sequences of alluvial sediments formed when the rate of sediment supply and the rate of creation of accommodation space are in equilibrium (Shanley and McCabe 1994). Changes in fluvial architecture can also result from climatic effects, such an increase in the amount, or distribution with time, of rainfall (Miall 1996), or a change in erosion or weathering processes. Units of FA1 are repeated throughout the Early Devonian succession as shown in figure 3.30, indicating that the physical and palaeoenvironmental conditions necessary for their deposition (Fig. 3.15) occurred a number of times.

The sedimentary rocks comprising FA1 are interpreted to have been deposited during a period of high and widespread accommodation space. This typically occurs when base level is rising and there is an upward shift in the graded fluvial profile (Posamentier and Vail 1990). Such periods of rapid base level rise occur in the TST and the early HST and consequently this facies association is interpreted as such (Fig. 3.30).

Facies Association 2

The fluvial architecture of FA2 can be used to interpret the base level conditions during their deposition. The channel fill sandstones in FA2 are truncated by numerous laterally continuous scour surfaces to form a multi-storey channel complex (Fig. 3.30) which can be traced laterally across many tens of kilometres. Units of FA2 form laterally continuous sandstone sheets within which examples of the other facies associations (FA1, FA3, FA4, FA5, and FA6) were not observed. Sedimentary rocks comprising FA2 occur a number of times during the Early Devonian indicating that the conditions controlling their deposition occurred a number of times.

The vertical frequency of scour surfaces and channel scour surfaces in the multi-storey sand bodies indicates a high degree of reworking of the previously deposited channel fill sand bodies. Such reworking is interpreted to indicate a disequilibrium in the alluvial graded profile which can result from fluctuating flow conditions (Miall 1996). The widespread nature and vertical repetition of the scour surfaces in FA2 indicate that low flow stage conditions dominated and that generally low amounts of accommodation space were present across a wide area. This regional distribution suggests that the factors influencing the alluvial system were not confined to localised geographic areas or active during brief periods of an alluvial discharge cycle, i.e. within incised valleys or during low stage conditions. Although the alluvial architecture indicates low amounts of accommodation space within the alluvial network, the laterally continuity of the sandstone bodies suggests that sediment supply and accommodation space were close to equilibrium.

The origin of the limited amounts of accommodation space, during the deposition of FA2 may be autocyclic or allocyclic in origin. The fluvial architecture, vertical repetition, and lateral continuity of FA2 are interpreted to result from repeated times when the rate of creation of accommodation space was low (Fig. 3.30). During a base level cycle (Fig. 2.8), times when the rate of creation of accommodation space is low, can occur in the latter stages of the HST and during the FSST or LST (Fig. 3.30). Units of FA2 are thus interpreted to have been deposited when base level conditions were at a stillstand, falling, or the rate of base level rise was low during the late HST, FSST or LST (Fig. 3.30).

3.9.2. *Sequence Stratigraphic interpretation for Facies Association 3.*

Facies Association 3 is interpreted as abandoned channel fill deposits in a braidplain system. According to Miall (1996) deposits such as these can form within fluvial systems as a result of channel switching or a reduction in coarse clastic sediment to the area. The laterally discontinuous nature of FA3 may be a primary feature or may represent the isolated preservation of a once widespread lithofacies (figures. 3.15 and 3.15c).

It is problematic to interpret FA3 in terms of accommodation space and base level. Facies association 3 mainly comprises the fill of abandoned channels, i.e. accommodation space that is already present. As such, FA3 do not provide any direct indication of the rate of creation of accommodation space during their deposition, rather their preservation indicates that an absolute amount of accommodation space was present. However, the ichnofacies in FA3 indicates that these units may represent the isolated remnants of once widespread paralic deposits. Therefore, the juxtaposition of the channel fill deposits of FA3 with the sandy braidplain deposits of FA1 and FA2 may indicate the transition from a braidplain to an alluvial flood plain or paralic setting (figures 3.15, 3.20 and 3.23). This transition may indicate that FA3 was deposited during or after a rise in base level, and are therefore tentatively interpreted as TST or HST deposits (Fig. 3.30).

3.9.3. *Sequence Stratigraphic interpretation for Facies Association 4.*

Facies Association 4 are interpreted as alluvial flood plain and coastal plain deposits (Fig. 3.20). In the absence of data from the associated fluvial or deltaic channel deposits it is difficult to interpret the transition to FA4 in terms of a change in fluvial style from braided to anastomosing or meandering channels. However, the increase in the amount of fine-grained material in FA4 indicates an increase in the amount and preservation of, fine-grained material relative to FA1 and FA2.

Changes in fluvial style can be associated with changes in discharge, bedload and overbank character, palaeoclimate, and the gradient of the fluvial system (Shanley and McCabe 1994; Miall 1996). During the Early Devonian the vegetation cover in continental areas was limited, being mainly confined to areas near the palaeoshorelines (Schumm 1969; Miall 1996). These authors also state that this lack of vegetation cover in interfluve areas to stabilise river banks will have promoted the braiding of alluvial systems. The palaeoclimatic conditions in the Murzuq Basin are interpreted to have resulted in alluvial systems similar to those found in modern arid regions (section 1.4.2, Fig. 1.13). According to Emery and Myers (1996) and Miall (1996), a change in fluvial channel pattern from braided to anastomosing or meandering represents a decrease in channel sand body connectivity which can be associated with a rise in relative sea level and an upward shift in the graded profile. Base level rise can result from a decrease in the gradient of the fluvial system, which may be due to aggradation related to deltaic progradation (Emery and Myers 1996). The profile for the Murzuq Basin outlined in section 1.4.1 means that it is more likely that the base level rise interpreted to be associated with FA4 is controlled by a rise in relative sea level. The lateral continuity of FA4 deposits suggests that whatever the factor(s) controlling the change in fluvial style, the effects were geographically widespread. But, the factors controlling changes in fluvial style were also of (relatively) limited duration as FA4 are encased within units of FA1 and FA2 (Fig. 3.30). The ichnofabric in FA4 also indicates a paralic influence on their deposition (Fig. 3.20). Units of FA4 are therefore interpreted to have been deposited during, or following, a period of base level rise, and are therefore interpreted as TST or early HST deposits (Fig. 3.30).

3.9.4. *Sequence Stratigraphic interpretation for Facies Associations 5 and 6.*

Facies Associations 5 and 6 are interpreted as shallow to marginal marine deposits (Fig. 3.23), and generally overlie units of FA1 and FA2 (Fig. 3.30). This transition from predominantly alluvial to marine influenced facies indicates that a period of base level rise must have occurred prior to the deposition of FA5 and FA6.

The internal bedform geometry and juxtaposition of facies in FA5 and FA6 do not provide any clear indication of the rate of creation of accommodation space during their deposition. At locality 10 the succession comprises lower shoreface deposits overlain by tidal channel and tidal flat deposits, recording a minor shallowing-upward of water depth. The lack of data from coeval sediments laterally or down-dip mean that the origin of this change in palaeo-water depth may be autocyclic or allocyclic. The inter-relationships between FA5 and FA6 in terms of relative sea level are also difficult to confidently interpret. Although FA5 contains comparatively less evidence of marine influence than FA6, the lack of data from coeval sediments mean that it is impossible to distinguish between any autocyclic or allocyclic processes that may have been active during their deposition.

The base level rise associated with a transition from the alluvial facies in FA1 and FA2 to the marine influenced facies in FA5 and FA6 can be observed in broadly coeval facies throughout the study region (Fig. 3.30). Previous studies concerned with the Lower Devonian succession on the SW margin of the Murzuq Basin by Klitzsch (1969), Beuf et al. (1971), Bellini and Massa (1980), Galecic (1984), Jakolovljevic (1984), Protic (1984), Clarke-Lowes (1985), and Selley (1997a) have also noted an increased marine influence in the upper part of the Lower Devonian succession. This abrupt base level rise is interpreted as a regionally extensive base level rise and transgression with units of FA5 and FA6 interpreted as late TST and HST deposits (Fig. 3.30).

3.9.5. *Sequence Stratigraphic interpretation for Facies Association 7.*

Facies Association 7 are interpreted as shallow marine to marginal marine sedimentary rocks, deposited when the rate of siliciclastic sediment supplied to the area was low (Fig. 3.28). The internal character of FA7 provides little evidence of absolute base level changes during its deposition. The dramatic reduction in siliciclastic sediment supply between FA5 and FA6 and FA7 represents a shut off of sediment supply to the area. Abrupt decreases in the amount of coarse clastic sediment supplied to basinal areas can occur when sediment source areas up-dip are transgressed (Emery and Myers 1996), or when detatched shoreface successions form during regression (J. Howell 1999, pers comm.). The thin, fining and coarsening-upward cycles in FA7 were not traced laterally during this study. However, previous studies of these sediments identified similar fining and coarsening-upward successions along the SW flank of the basin (Galecic 1984; Jakolovljevic 1984; Protic 1984; Clarke-Lowes 1985). These thin fining and coarsening-upward cycles may be autocyclic in origin or alternately parasequences, representing periods of progradation. Due to limited data the factors driving their formation are poorly defined.

Units of FA7 are observed immediately overlying the siliciclastic marine influenced deposits of FA5 and FA6 (Fig. 3.30), representing an abrupt change in facies across the contact. The change in palaoewater depth between FA5, FA6, and FA7 is difficult to ascertain and may be minimal. The abrupt facies change across the contact can be used to interpret a major change in basin physiography and sediment supply processes to the shallow marine depo-systems. The presence of brackish and acidic lagoonal palaeoenvironments during the deposition of FA7 is interpreted by Galecic (1984), Jakolovljevic (1984), and Protic (1984) to indicate the presence of one or more closed basins that may indicate a tectonically driven change in basin physiography, following which depositional processes varied. The erosion of the Ouan Kasa Formation from the central and NW part of the western margin of the basin recognised by D. Massa (1998, pers comm.) indicates that the area was subject to tectonism during this time. The abrupt change in facies observed between the Tadrart and Ouan Kasa formations may indicate another phase of tectonism prior to the deposition of FA7, as well as an event that post-dates the deposition of FA7. Alternately the origin of this major change in facies may be autocyclic, e.g. a lateral shift in facies belts, or a change in sediment supply processes, or allocyclic e.g. changes in palaeoclimate or eustasy.

Examples of FA7 are recognised along much of the entire western margin of the Murzuq Basin, and therefore the abrupt change in facies is of regional importance, rather than localised within a sub-basin. According to the above factors the sedimentary rocks comprising FA7 are interpreted to have been deposited during the TST and HST of a relative sea level cycle associated with a major change the basin morphology. Alternately, sedimentary rocks comprising FA7 may represent a LST detached shoreface succession with the intervening TST and HST deposits absent. The base level conditions responsible for the deposition of FA7 are therefore difficult to ascertain from these data.

3.9.6. *Sequence boundaries in the Lower Devonian succession.*

The preservation of a number of intervals of Early Devonian age facies that are characterised by conditions of high and low accommodation space implies the presence of a number of sequence boundaries. The low angle basin profile and low rate of subsidence for the Murzuq Basin outlined in section 1.4.1 suggests that these will be type 1 sequence boundaries. These type 1 sequence boundaries will have formed during a relative sea level cycle within which conditions of relative sea level fall are extended relative to periods of relative sea level rise (Fig 2.13). The character of the systems tracts and key surfaces within the sequences are discussed in section 3.9.7.

The lower contact of the Lower Devonian succession is recognised as an unconformity surface, and therefore sequence boundary, throughout the Murzuq Basin according to Bellini and Massa (1980) and Clarke-Lowes (1985). A discussion relating to the possible origins of this lower bounding surface can be found in sections 1.4.1 and 6.5.1.

As well as the sequence boundary at the base of the Lower Devonian succession a number of candidate sequence boundary surfaces were identified from field studies of the facies associations previously outlined. The uncertainties associated with the sequence stratigraphic interpretation of FA2 makes the positioning of the many of the sequence boundary surfaces in the Lower Devonian Tadrart Formation problematic. The problematic identification of the sequence boundaries in the Lower Devonian succession also raises the question of the location of any flooding surfaces within the sequences. A number of iron-stained and pitted surfaces were observed during field studies within and between the facies associations outlined above (Figures 3.1 and 3.30). Most of these surfaces are of limited lateral extent and are interpreted to have formed as a result of autocyclic processes active within the fluvial and shallow marine palaeoenvironments. However, six of these iron- stained surfaces had a different physical character, regional extent, and represented surfaces across which there was a sharp change in facies (Fig. 3.30). According to these properties, these six surfaces can be interpreted as sequence boundaries, formed during periods of prolonged exposure, minor erosion and sediment by-pass during falls in base level, flooding surfaces associated with relative sea level rise, or a composite of the two key surface types. Previous work by Clarke-Lowes (1985) identified three sequence boundaries in the Lower Devonian succession on the western margin of the Murzuq Basin (Fig. 3.31) and surfaces with a similar character are also identified as sequence boundaries in the Lower Palaeozoic succession of Oman (Millson et al. 1996). However, the identification of one such ferruginous surface by Millson et al. (1996) as a sequence boundary in Oman is not accepted by Buckley and Harbury (1996), highlighting the problems with the identification of such sequence boundaries.

Due to the uncertainties associated with the sequence stratigraphic interpretations of some of the facies in the Lower Devonian Tadrart Formation two alternative sequence stratigraphic models have been constructed for these sedimentary rocks and can be seen in figure 3.32. Both sequence stratigraphic models identify six sequence boundaries in the Lower Devonian succession, the lower five (SB1 to SB5) between units of facies associations 1 to 6 (Tadrart Formation) (Fig. 3.32), and the sixth identified at the base of FA7 (Ouan Kasa Formation) (Fig. 3.32). In model a) units of FA2 are interpreted as late HST to early TST deposits commonly making the sequence boundary surfaces combined flooding surfaces (Fig. 3.32). In model b) units in FA2 are interpreted as LST deposits. The differing sequence stratigraphic interpretations of FA2 within models a) and b) reposition sequence boundaries one to five as well as the flooding surfaces associated with these relative sea level fluctuations (Fig. 3.32). Using model a) the laterally continuous ferruginous-stained surfaces are sequence boundaries while in model b) their importance is diminished as they occur within sequences as flooding surfaces and systems tract boundaries, as well as marking sequence boundaries (Fig. 3.32).

The position of the two upper sequence boundaries (**SB5** and **SB6**) are well defined and consistent in both models, with **SB5** separating the braidplain facies of FA1 or FA2 and the marginal to shallow marine facies in FA5 or FA6 throughout the study region (Fig. 3.32). The uppermost sequence boundary in the Lower Devonian succession is sequence Boundary six (**SB6**), being heavily pitted, irregular and iron-stained at localities 4, 11, and 12. This sequence boundary surface marks the contact between the siliciclastic shallow marine deposits of FA5 or FA6 and the fine-grained oolitic facies in FA7. Sequence boundary six (**SB6**) therefore represents a surface across which there is a pronounced change in depositional style, regardless of whether the sedimentary rocks in FA7 are interpreted as TST to HST deposits (model a) or LST deposits (model b). As the character of **SB6** is different to the underlying sequence boundaries it may have formed when palaeoclimatic conditions were different to those that were active when previous sequence boundaries were formed. Alternately, **SB6** may have formed when the basin physiography and resultant depositional processes were altered by a phase of tectonism and/or a fall in relative sea level during the formation of **SB6**. If FA7 represents a detached lowstand shoreface then it is highly unlikely that it is associated with the base level cycle responsible for the deposition of the underlying HST deposits in FA5 and FA6. The palaeowater depth during the deposition of FA5 and FA6 is approximately the same or shallower than that interpreted during the deposition of FA7. Therefore FA7 cannot have been deposited following a single fall in relative sea level but may comprise the LST of a successive cycle with units from the intervening TST, HST and FFST absent from the entire region.

Although both models 'work' in terms of a sequence stratigraphic interpretation for the Lower Devonian succession the sequence stratigraphic interpretations in model a) are slightly favoured to those made in model b). This is primarily because of the basin physiography and low rates of subsidence of the region during the Lower Devonian. Such conditions make the preservation of thick, laterally continuous LST deposits unlikely and therefore it is more probable that the deposits comprising FA2 were deposited during late HST to FSST's, although minor amounts of erosional re-working and deposition cannot be entirely discounted during subsequent LST's. The relative sea level fluctuations identified in model a) are discussed below, in terms of the systems tracts, sequences and key surfaces in the succession.

3.9.7. *Lower Devonian base level/relative sea level fluctuations, systems tracts, sequences, and key surfaces.*

Using the interpretations in sequence stratigraphic model a) the six sequence boundaries within the Lower Devonian succession separate seven depositional sequences (S1 to S7) (Fig. 3.32) which contain one or more of the facies associations outlined above. The lower six sequences are identified in the lithostratigraphic unit corresponding to the Tadrart Formation while the uppermost sequence identified corresponds with the Ouan Kasa Formation.

As stated above, using sequence stratigraphic model a) there is a general absence of abundant sedimentary rocks deposited during base level lowstands (LST deposits). This absence may be due to their attenuated thickness, limited lateral distribution, and the regional nature of this study. However, the slowly subsiding, low angle ramp type profile of the Murzuq Basin outlined in sections 1.4.1 and 2.4 will have inhibited the deposition and preservation of LST deposits in the Murzuq Basin. During falls in relative sea level the Murzuq Basin may have predominantly been an area of minor erosion or sediment by-pass with the palaeoshoreline dislocated many tens or hundreds of kilometres down-dip as outlined in figure 2.12. One such detached lowstand shoreline may be represented by the sedimentary rocks comprising FA7. The low angle basin slope and low rate of subsidence within the basin during this time will have also have lengthened the duration of any periods of relative sea level fall in the basin (Fig. 2.13). Consequently periods of falling or low base level will dominate the stratigraphic record within the Murzuq Basin during the Early Devonian.

Base level cycles or relative sea level fluctuations usually result in the deposition of a number of systems tracts that are separated by key sequence stratigraphic surfaces (appendix 2.1). The sequence boundaries outlined in section 3.9.6 and figure 3.32 separate such intervals containing TST, HST, FSST and possibly LST deposits (Fig. 3.32). Therefore the sequence boundaries juxtapose sediments that were deposited when variable amounts of accommodation space were present. Using model a) the transition from conditions of low to high(er) accommodation space, across the sequence boundaries outlined above and shown in figure 3.32, indicates a phase of relative sea level rise associated with a flooding surface. Therefore in the case of the Murzuq Basin these sequence boundary surfaces are also combined flooding surfaces as LST deposits are thought to be almost entirely absent (see above). These flooding surfaces cannot be sub-divided in terms of initial and maximum (see appendix 2.1) because of the limited data relating to their relationship with the underlying strata and facies up-and down-dip.

A typical Early Devonian base level cycle resulted in the deposition of facies associations characterised by high or rising base level, such as FA1, FA3 or FA4, overlain by FA2 deposited during conditions of falling or low base level. This juxtaposition illustrates a change from deposition during TST's through to FSST's on a base level cycle (Fig. 2.8), in this instance responsible for changes in alluvial architecture (section 3.2.3). Once the rate of relative sea level fall exceeded a critical point, the latter factor dependant on sediment supply and the basin profile, the preservation of alluvial sediments throughout much of the region ceased and sequence boundary formation will have begun.

For deposition of sequence one (S1) to have occurred after the Late Silurian/ Early Devonian tectonism and regional relative sea level fall outlined in section 1.4.1, accommodation space needs to have been created. The lower part of S1 comprises units of FA1 which indicates that moderate to high amounts of accommodation space were available. The upper part of S1 comprises a succession of FA2. Sequence one is therefore interpreted to have been deposited as a result of a rise in rate of creation of accommodation space followed by a decrease in the rate of rise (figures. 3.32 and 3.33).

Sequence two (S2) comprises units of FA3 at its base which are overlain by sedimentary rocks of FA1. This juxtaposition indicates an initial moderate rate of creation of accommodation space which is followed by a decrease in the rate of rise (Fig. 3.32). The decrease in the rate of creation of accommodation space in S2, is interpreted to be the result of the progradation of HST fluvial systems following the TST (figures. 3.32 and 3.33).

Sequence three (S3) comprises units of FA4 at its base, which are overlain by deposits of FA1 and then FA2 (Fig. 3.32). Although the degree of confidence associated with the sequence stratigraphic interpretation of FA4 is limited, the vertical juxtaposition of facies in S3 is interpreted to indicate an initial high rate of creation of accommodation space followed by a decrease in the rate of rise or a fall. (figures 3.32 and 3.33). This reduction in the rate of creation of accommodation space is interpreted to be the result of prograding HST fluvial systems and/or a decrease in the rate of base level rise (Fig. 3.33). The transition from FA1 to FA2 in the upper part of S3 is interpreted as a change in fluvial style, driven by changes in the rate of creation of accommodation space that regraded the alluvial profile. As outlined by Shanley and McCabe (1993, 1994), Emery and Myers (1996) and Miall (1996) changes in fluvial architecture can be used to interpret changes in the rate of creation of accommodation space.

Sequences four and five (S4 and S5) comprise a succession of FA1 overlain by deposits comprising FA2. The facies juxtaposition in S4 and S5 is interpreted to indicate an initial increase in the rate of creation of accommodation space following sequence boundaries four and five (SB4 and SB5), succeeded by a decrease in the rate of rise or a slight fall (Fig. 3.32). In accordance to the sequence stratigraphic interpretations assigned to FA1 and FA2 the two sequences were deposited during the TST, HST, and FSST's as outlined above (figures 3.32 and 3.33).

Sequence six (S6) comprises a succession of sedimentary rocks interpreted as FA5 and FA6 which immediately overlie units of FA1 and FA2, separated by a sequence boundary (SB5) (Fig. 3.32). The change from the non-marine units of FA1 and FA2 to FA5 and FA6 indicates a rise in relative sea level following SB5 (Fig. 3.32). Facies Associations 5 and FA6 in S5 were deposited during and following a rise in relative sea level and are placed within the TST and HST (Fig. 3.32). The minor changes in palaeowater depth associated with the changes in and between FA5 and FA6 may be the result of autocyclic or allocyclic controls. The limited data available on the inter-relationships and regional distribution of FA5 and FA6 mean it is impossible to interpret the driving mechanism(s) of these facies changes, i.e. whether these are autocyclic or allocyclic high frequency cycles in relative sea level.

Sequence seven (S7) comprises deposits of FA7 which are separated from the overlying sedimentary rocks of FA5 and FA6 by a sequence boundary (SB6) (figures 3.32 and 3.33). The abrupt changes in depositional style that occur across this surface are interpreted to result from a basin wide change in the physiography of the basin, related to the formation of SB6 (chapter 6).

The sequence stratigraphic interpretations for facies associations 1 to 7, their juxtaposition outlined above, and the position of sequence boundaries one to six can be used to construct a Lower Devonian relative sea level curve (Fig. 3.33). This relative sea level curve is constructed in terms of the rate of creation of accommodation space that is predominantly controlled by subsidence and eustasy (section 2.4.2).

3.9.8. *Interpreting the origin and distribution of relative sea level fluctuations in the Lower Devonian succession on the SW margin of the Murzuq Basin.*

The origins of the fluctuations in relative sea level outlined above need to be considered to further understand the history of the basin. The relative sea level curve constructed for the Lower Devonian comprises seven sequences (Figures 3.32 and 3.33).

The lowermost six sequences identified (S1 to S6) occur in the Tadrart Formation which is dated Gedinian to early Emsian (Selley 1997a), approximately spanning 408.5 Ma to 386 Ma respectively according to Harland et al. (1989). The six sequences were therefore deposited over a maximum of 22.5 Myr (Fig. 3.33): which would give an average of 3.75 Myr per sequence if each sequence had the same duration. The duration of each sequence also includes any periods of non-deposition and/or erosion during the base level cycle. As the upper part of the Tadrart Formation includes lower Emsian sedimentary rocks in S6 it is impossible to estimate the position of the lower Emsian boundary in S6 and therefore the duration of all of the Lower Devonian sequences. To simplify correlation it is interpreted that the upper part of S6 continues for approximately 2.5 Myr into the Emsian, terminating at 388 Ma. Although the Ouan Kasa Formation (S7), which is dated Emsian to lower Eifelian (Fig. 3.33), was deposited over a maximum of 12 Myr (390.5 Ma to 384 Ma), it is unlikely that the deposition occurred for this entire time.

For the purposes of this study the duration of each of the Lower Devonian sequences is interpreted to be approximately 3.5 Myr which would make these comparable to the duration of the 3rd order cycles outlined by Emery and Myers (1996) (Fig. 2.6). These cycles of 3rd order duration are interpreted to have strongly influenced alluvial architecture in the lower five sequences (S1 to S5, Fig. 3.32) and the juxtaposition of shallow marine facies in the upper two sequences (S6 and S7, (figures 3.32 and 3.33).

Sequences one to seven record an overall rise in relative sea level punctuated by higher frequency, lower magnitude fluctuations interpreted as 3rd order cycles (Fig. 3.33). This relative sea level rise is clearly seen by the increased amounts of marine influence observed in the upper part of the Lower Devonian succession. The origin of the overall rise in relative sea level recognised in the Lower Devonian succession may be due to a 2nd order eustatic cycle superimposed upon the 3rd order cycles, or the localised effects of an increase in the rate of subsidence. The potential regional correlation of the Early Devonian relative sea level fluctuations outlined above are discussed in chapter 6.

It is recognised that the dating of the Lower Devonian sedimentary rocks and base level cycles contained therein is only an approximation not an absolute figure because of the poor biostratigraphic data. It is reasonable to assume that active deposition within these sequences only occurred during part of each base level cycle, possibly as low as 50 % for the Murzuq Basin (Fig. 3.33) (section 2.4.1).

3.9.9. *The effects of relative sea level fluctuations during the Early Devonian.*

The driving mechanisms responsible for the relative sea level fluctuations outlined above and the resultant evolution of facies during the Early Devonian need to be considered. As detailed in chapter 2 and by Shanley and McCabe (1993, 1994) in an up-dip area, the effects of eustasy decrease relative to the effects of climate, tectonism, and the autocyclic processes active within the deposystems (Shanley and McCabe 1993; 1994; Emery and Myers 1996; Miall 1996). For example, according to Shanley and McCabe (1994), Emery and Myers (1996), and Howell and Davies (1996), within the modern Mississippi fluvial system the effects of recent eustatic fluctuations are only noted 150 to 200 kilometres inland. The position of the palaeoshoreline (and bayline) during the Early Devonian is very difficult to establish during much of this time due to a lack of data from coeval facies down-dip. As a result, the degree of confidence associated with the sequence stratigraphic interpretations for the alluvial facies (FA1, FA2, FA3 and FA4) is difficult to quantify. However, the identification of a number of unconformably bound sequences containing different alluvial and shallow marine facies suggests the controlling mechanisms are cyclical, and strongly influence facies development. The major controls of climate, tectonism and eustasy strongly influence facies development and their juxtaposition, with a change in one of these parameters likely to be reflected in the others as they are somewhat inter-dependent (Shanley and McCabe 1994).

According to Emery and Myers (1996) 3rd order cycles such as those interpreted from this study in the Lower Devonian succession are generally thought to be of glacioeustatic origin. However, it is proposed by Cloetingh (1988), Emery and Myers (1996) and Miall (1996; 1997) that cyclic variations in climate and tectonism may also be responsible for the generation of 3rd order cycles.

Shorter time scale fluctuations in the climate within the Murzuq Basin are difficult to establish due to a lack of conclusive data from fauna, palaeosols, and also data from lateral and down-dip areas. However, studies on alluvial successions, from a number of stratigraphic intervals throughout the world, have identified cycles in alluvial architecture which appear to be strongly influenced by variations in palaeoclimate (Lawrence and Wiliams 1987; Shanley and McCabe 1994; Kelly and Sadler 1995; McKie and Garden 1996; Miall 1996, 1997). The variations in facies developed during the Early Devonian may be in part caused by climatically induced fluctuations in alluvial discharge and bedload character. However, it is unlikely that these factors had any influence on the overall rise in relative sea level recognised from this study in the Lower Devonian succession in the Murzuq Basin.

The development of, and juxtaposition, of facies associations in the Lower Devonian succession occurred as a result of variations in base level/relative sea level. These fluctuations lead to changes in the rate of creation of accommodation space, resulting in the modification of the fluvial graded profile, the juxtaposition of fluvial and shallow marine facies, and the formation of sequence boundaries. It is possible that these fluctuations resulted in slight changes in the palaeoclimate and the erosional and weathering processes associated with varying climatic conditions. The possible driving mechanisms of the cycles identified within the Lower Devonian succession on the western flank of the Murzuq Basin are discussed in chapters 2 and 6.

Chapter 4. Outcrop log correlation and interpretation for the Middle and Upper Devonian succession of the Murzuk Basin, southern Libya.

This chapter contains data and interpretations relating to Middle to Upper Devonian sedimentary rocks that were studied on the northern margin of the Murzuq Basin (Fig. 1.3). Previous work in the area undertaken by Seidl and Rohlich (1984) subdivided the Middle to Upper Devonian succession into six formations (Fig. 1.4), the B'ir al Qasr, Idri, Quttah, Dabdab, Tarut and Ashkidah formations. The stratigraphic ages of the formations shown in figure 1.4 were established by previous authors including Seidl and Rohlich (1984), Parizek et al. (1984), Massa (1988) and Mergl and Massa (1992).

This study has reinterpreted the Middle to Upper Devonian succession in terms of their environments of deposition and, due to the nature of the dataset, their sequence stratigraphic evolution. Therefore, the Middle to Upper Devonian succession has been subdivided into seven parasequence sets (PS1 to PS7) which contain numerous facies types. To avoid repetition, each allostratigraphic unit, i.e. PS1, is detailed and interpreted in turn, rather than listing the facies associations.

4.1. Bir' al Qasr Formation (Eifelian to Lower Givetian).

The B'ir al Qasr Formation crops out on the southern flank of the Gargaf Arch and unconformably overlies sedimentary rocks of the Tanezzuft Formation (Silurian) in the west of the study region (Fig. 4.1) and undifferentiated Cambro-Ordovician sandstones in the central and eastern parts of the study area (Fig. 4.2a). Two main units can be identified within the B'ir al Qasr Formation (figures 4.2b and 4.3): a lower parasequence set 1 (PS1), and an upper parasequence set 2 (PS2). Parasequence set 1 contains a maximum of a basal lag and up to three parasequences (Fig. 4.3), while parasequence set 2 contains a maximum of three parasequences (Fig. 4.3). The correlation of key surfaces in PS1 and PS2 between areas is problematical due to incomplete and poorly exposed sections through the B'ir al Qasr Formation and facies variations along strike and up dip. A graphic log correlation panel of the B'ir al Qasr Formation can be seen in figure 4.4, which also displays palaeocurrent data from the logged sections and detailed study areas and the position of these locations. A colour coded basin profile has been constructed for deposition of **PS1** and **PS2** (Fig. 4.5) upon which the migration of facies belts in these stratigraphic units can be superimposed. A list of detailed study areas, logged sections and their locality numbers can be found in appendix 4.1. In the following text, the data and their interpretations are listed to form a west to east transect of the outcrop in an up-dip and along-strike traverse.

4.1.1. *Parasequence set 1 (PS1). Basal lag.*

The lowermost deposits of **PS1** form a basal lag which is compositionally different from the overlying sedimentary rocks. The basal lag is observed at the base of the Middle Devonian succession in the central and western parts of the study area (Fig. 4.6). At locality 21 (log 21) the basal lag is 1 metre thick and comprises a poorly sorted medium to coarse-grained sandstone with intraclasts of ferruginous sandstone (Fig. 4.6), mudstone flakes, ferruginous-stained shell moulds, bone fragments and rare *Planolites* or *Tigillites* burrows. Parizek et al. (1984) also observed fragmentary fossils and occasional *Tigillites* burrows in the lag deposit at this locality. A basal lag is also observed at locality 25 (log 23) and possibly locality 4 (log 24). At locality 23 the basal lag comprises 80 centimetres of ferruginous sandstone containing weathered-out mudstone clasts. At locality 24 the basal lag may be up to 6 metres thick. However, poor exposure of the succession prevented confirmation of their identification.

The composition of the basal lag indicates that it is related to the overlying lowermost Middle Devonian age sedimentary rocks in its grain-size, clast population and faunal content rather than the Lower Devonian Tadrart and Ouan Kasa formations observed in the southwest of the Murzuk Basin and detailed in chapter 3.

4.1.2. *Parasequence set 1 (PS1). General description.*

At locality 21 parasequence 1 (**P1**) comprises a succession of moderately to poorly sorted medium-grained sandstones (Fig. 4.7) which contain crinoid fragments

(Fig. 4.7) and shell moulds. The ichnofabric in the sandstones is of high abundance (Fig. 4.8), comprising *Thalassinoides, Chondrites, Tigillites, Taenidiun Camerorensis, Phoebichnus Trachoides, Skolithos Ramosus* and *Planolites* (D. McIlroy 1998, pers comm.), with common lateral and vertical variations. At locality 25 only the lower 7 metres of **P1** are observed, comprising lenticular and tabular bedded siltstones and sandstones (Fig. 4.9). The sandstones contain asymmetrical and symmetrical ripples, planar tabular and trough bedforms, numerous internal erosion surfaces (including low angle truncation surfaces), erosively-based channels and low angle planar laminated sandstones. The sandstones also contain numerous flakes and rounded clasts of mudstone that are commonly weathered-out. Some of the sandstones also contain an ichnofabric of low to high abundance, comprising *Planolites, Thalassinoides, Tigillites* and *Chondrites,* with lateral variation in their diversity and abundance commonly observed. At locality 24 **P1** comprises iron-stained medium sandstones which are poorly exposed. Parasequence 1 was also studied in detail at locations 26, 27 and 29. At these localities **P1** comprises interbedded mudstones, siltstones, fine-grained sandstones and ferruginous medium to coarse-grained sandstones. The sandstones contain numerous siltstone and mudstone clasts, small erosive channels, shell moulds, fragments of plant imprints, symmetrical and asymmetrical ripples and an ichnofabric of low to high diversity and abundance. At locality 36 (log 25) **P1** comprises interbedded sandstones, siltstones and mudstones. The upper beds of the sandstones are iron-stained, pitted and contain symmetrical ripples. Internally the sandstones contain weathered-out clasts, mudstone clasts and an ichnofabric of low abundance comprising *Planolites* and *Tigillites.* The siltstones are purple, micaceous and contain an ichnofabric of medium abundance comprising *Planolites, Tigillites,* and *Chondrites.* At locality 31 (log 26) **P1** was not studied in detail. However, the interbedded sandstones, siltstones and mudstones beneath the base of log 26 are similar to those observed in **P1** at locality 36.

At locality 21 parasequence 2 (**P2**) comprises heterolithic siltstones and sandstones with interbeds of sandstone (Fig. 4.10). The heterolithic units contain an ichnofabric of medium to high abundance, comprising *Planolites, Tigillites,* *Thalassinoides* and *Chondrites*, while the sandstones contain an ichnofabric of low to high abundance comprising *Planolites*, *Tigillites*, *Skolithos*, and *Thalassinoides*. The sandstones form inclined bedsets (Fig. 4.10), contain shell moulds (Fig. 4.11), and the upper beds of the bedded sandstones contain symmetrical ripples. The succession terminates with a sandstone-filled channel containing a monotype *Skolithos* ichnofabric. At locality 24 **P2** comprises a coarsening-upward mudstone to very fine-grained sandstone/siltstone package. The siltstones contain an ichnofabric of low to medium abundance comprising *Planolites*, *Tigillites*, and *Spirophyton*, (Fig. 4.12) while the ichnofabric in the sandstones is of medium to high abundance, comprising *Planolites*, *Tigillites*, *Diplocraterion* and *Thalassinoides*. The units of **P2** contain symmetrical ripples with the upper surface pitted. At localities 22 (log 22) and 23 **P2** is poorly exposed, comprising heterolithic siltstones and shales, thin sandstones and shales. An ichnofabric is rarely observed, restricted to one of low to medium abundance comprising *Tigillites* and *Planolites* when observed. At locality 36 **P2** comprises a thick succession of varicoloured shales with occasional thin lenses and beds of siltstone. The shale appears barren of any body or trace fossils while the siltstone beds contain a low abundance *Planolites* and *Tigillites* ichnofabric. At locality 31 **P2** comprises a succession of shales with rare laterally discontinuous very fine sandstone beds which are overlain by sharp based siltstones and very fine-grained sandstones. These deposits are incised by sandstone-filled channels and sandstone bodies containing epsilon cross-bedding (Fig. 4.13). The lowermost studied shales of **P2** at locality 31 contain an ichnofabric of low diversity and abundance, while the overlying sharp based siltstones and very fine-grained sandstones contain a high abundance, generally a monotype, *Skolithos* or *Chondrites* ichnofabric. The epsilon cross-bedded and channelled sandstones contain an ichnofabric of low to medium diversity comprising *Planolites*, *Rhrizocorralium* and *Thalassinoides*. Fine-grained sediments that occur laterally to the epsilon cross-bedded sandstones contain a restricted ichnofabric of *Skolithos* only.

At locality 21 parasequence 3 (**P3**) comprises interbedded sandstones and siltstones. The sandstones contain minor surfaces of erosion, occasional mudstone clasts, symmetrical ripples, and an ichnofabric of low to medium abundance comprising *Planolites* and *Tigillites*. The siltstones are micaceous, fissile, occasionally heterolithic with very fine sandstone, and contain an ichnofabric of medium abundance comprising *Planolites*, *Chondrites* and *Thalassinoides*. At locality 24 **P3** comprises a coarsening-upward mudstone to very fine-grained sandstone and siltstone package. The siltstones contain an ichnofabric of low to medium abundance comprising *Planolites*, *Tigillites* and *Spirophyton*, while the ichnofabric in the sandstones is of medium abundance comprising *Planolites* and *Thalassinoides*. The upper beds of **P2** contain symmetrical ripples with the upper surface being iron-stained and pitted. At locality 22 **P3** comprises lenticular bedded sandstones with minor scour surfaces; they also contain an ichnofabric of low abundance comprising *Planolites* and *Tigillites* as well as plant debris. At locality 21 **P3** comprises interbedded siltstones and bioturbated sandstones. The sandstones are thinly bedded, erosively-based and contain abundant plant debris and an ichnofabric of low to medium abundance comprising *Thalassinoides*, *Skolithos*, *Tigillites*, *Planolites* and *Chondrites*. The sandstones also contain planar tabular and trough cross-bedding, low angle planar laminae, symmetrical and lunate ripples; several of the sandstone beds are cemented by iron. The siltstones are micaceous and they do not contain a clear ichnofabric. Several horizons contain vertical tubes that are interpreted as rootlets, although these may be vertical burrows (*Tigillites*). At locality 36 **P3** comprises poorly exposed grey/green/ brown shale incised by a 5 metre thick succession of sandstones with rare siltstone interbeds. The sandstones are very fine to medium-grained and generally well sorted although the uppermost sandstones are poorly sorted. The sandstones contain numerous erosion surfaces, soft sediment deformation structures, planar tabular and trough cross-bedding, low angle planar laminae, asymmetrical ripples, epsilon-cross bedding and plant debris. These sandstones also contain an ichnofabric of low abundance comprising *Planolites*, *Tigillites* and *Rhizocorallium* which is usually observed on bedding or bedset bounding surfaces. The siltstone interbeds are red, micaceous, fissile and contain occasional medium-grain size clasts of quartz sandstone. The uppermost sandstones of **P3** at locality 36 contain low angle planar laminae, mudstone flakes and an ichnofabric of low abundance comprising *Planolites* and *Tigillites* with the upper surface heavily pitted.

4.1.3. *Palaeocurrent data from PS1.*

Palaeocurrent data were collected from the logged sections and detailed study areas. These data are displayed in figure 4.14. However, the comparison of these data between outcrops is problematic due to lateral and vertical facies variations. Palaeocurrent data measurements included planar tabular and trough cross-bedding, low angle planar laminae, various ripple forms (symmetrical, asymmetrical and lunate), epsilon cross-bedding and inclined bedset bounding surfaces.

The variability observed in the planar tabular data is not unexpected as planar tabular bedform foresets can strike, and therefore migrate, anywhere between normal and parallel to the flow direction according to Selley (1988).

4.1.4. *Palaeoenvironmental interpretation for PS1.*

The basal lag and parasequences 1, 2, and 3 in PS1 were deposited in open marine to fluvial deltaic conditions. An interpretation of the palaeoenvironmental evolution of the area during this time can be seen in figure 4.15.

The basal lag formed during the early Mid Devonian, possibly the result of erosion and reworking of older rocks. The depositional environment of the basal lag is difficult to ascertain. However, the clast content and ichnofabric indicate that shallow to marginal conditions prevailed. Similar palaeoenvironmental interpretations were reached by Parizek et al. (1984) who interpreted the presence of the *Tigillites* ichnofabric to indicate a marginal marine setting. The basal lag thickness and occurrence of fragmentary fossils, bone debris and iron-stained shell material indicate that continental and/or marine sediments which had been deposited during the Lower or early Middle Devonian, were re-worked during the early stages of a slow phase of relative sea level rise (transgression) during the early Middle Devonian (section 4.7). There is no lithological or faunal evidence that the basal lag at locality 21 is the Akakus Formation (Silurian) as proposed by Blanpied and Rubino (1998). The basal lag is absent from the east of the study area at localities 31 and 36 (Fig. 4.4) where **P1** immediately onlaps Cambro-Ordovician sandstones (Fig. 4.2a), a relationship also noted by Seidl and Rohlich (1984). The onlap of the lower Middle Devonian sedimentary rocks onto Cambro-Ordovician sediments on the southern flank of the Gargaf arch was also noted on the southeastern flank of the Murzuq Basin by Bellini and Massa (1980), and also in the Djado sub-basin to the south by Mergl and Massa (1998). The limited geographic extent of the basal lag on the southern flank of the Gargaf arch is interpreted to be the result of localised variations in accommodation space formed by the pre-existing palaeotopography and post-depositional erosion. The influence of a palaeotopography as a control on the distribution and thickness of the B'ir al Qasr Formation on the southern margin of the Gargaf Arch was also recognised by Seidl and Rohlich (1984).

The basal lag or pre-Middle Devonian unconformity surface is succeeded by parasequence 1 (**P1**) which comprise shallow marine sediments throughout the study area, corresponding to the palaeoenvironmental interpretations of Mergl and Massa (1992). These interpretations indicate a rise in relative sea level and the creation of region-wide accommodation space. The distribution of palaeocurrent data suggests that the regional dip was towards the W or N with the palaeoshoreline oriented N-S to NNE-SSW during this time. Sedimentary rocks comprising **P1** at locality 21 are interpreted as shallow marine facies deposited in the offshore-transition zone, which become increasingly influenced over time by delta front and tidal processes. At localities 24 and 25 the frequent occurrence of low angle planar laminated sandstones and low angle truncation surfaces in **P1** indicate a high energy environment of deposition. The diverse and abundant marine ichnofabric in **P1** at localities 24 and 25 indicate shallow marine conditions. These sedimentary rocks also contain herringbone cross-bedded sandstones, formed by shoreline perpendicular N/S tidal currents. These tidally generated bedforms support a shallow marine palaeoenvironment during the deposition of **P1** locality 25. The sedimentological and biogenic features observed in **P1** at localities 26, 27, 29 and 36 indicate a shallow to marginal marine palaeoenvironment in the east of the study area. The extensive asymmetrical,

symmetrical and interference ripples and lenticular bedded sandstones in **P1** at localities 26, 27, 29 and 36 can be characteristic of deposition in a tidal or intertidal flat environment (Elliott 1986). The distribution of planar and trough data from locality 27 would indicate that although ebb directed tidal currents dominate, NW/SE oscillatory currents were also acting to drive the migration of planar and trough bedforms. The distribution of planar bedform current data suggests that oscillatory wave action may have been the dominant mechanism driving planar bedform migration while ebb directed currents were the primary driving mechanism of trough bedform migration. Moderate to normal current activity is also interpreted during the deposition of these sediments by Mergl and Massa (1992). The palaeocurrent data from locality 27 is interpreted to indicate the preservation of trough bedforms deposited within the main ebb channel and planar tabular bedforms deposited within the flood channels situated marginal to the ebb channel. These bedform relationships are also recognised in tidal deltas by Imperato et al. (1988). At localities 31 and 36 **P1** was deposited in tidal/intertidal mudstone flats which were prone to subaerial exposure. The end of **P1** is characterised by a regionally traceable surface across which there is a sharp change in facies and grain-size. This is interpreted as a parasequence boundary, representing a surface across which there is an increase in relative sea level.

Parasequence 1 (**P1**) is overlain by parasequence 2 (**P2**). At locality 21 **P2** was deposited within the offshore transition zone, shoreface and delta front. The latter palaeoenvironment was also recognised by Vos (1981); De Castro et al. (1991); El-Rweimi (1991); and Pierobon (1991). In the central and eastern part of the study area **P2** comprises fine-grained deposits, in the west of the study area these are interbedded with sandstones. The lower part of **P2** records a major decrease in mean grain-size throughout most the area. The diverse marine ichnofabric and sedimentary structures indicate the deposition of **P2** at locality 24 in a marine environment that may have been anywhere between a low energy lower shoreface or offshore transition zone. At locality 22 **P2** is interpreted to have been deposited in a low energy flood plain, delta front or interdistributary bay environment. Deposition of **P2** at localities 31 and 36 is interpreted to have occurred within the lower shoreface to offshore transition zone which is overlain by a tidal flat cut by laterally accreting tidal channels giving epsilon cross-bedding. The restricted ichnofabric observed in the lower part of **P2** may indicate adverse palaeoenvironmental conditions such as a freshwater influx to the area or anoxia. Alternately, there may be a highly diverse and abundant ichnofabric in the lower part of **P2** which is unclear. The fine-grained sedimentary rocks in the lower part of **P2** are overlain and incised into by siltstones and sandstones to give a vertical coarsening-upward cycle. This cycle contains sedimentary structures and ichnofacies indicating an upward shallowing of water depth. The thickness of the fine-grained deposits in the lower part of **P2** indicate a prolonged period when low energy conditions prevailed and clastic input was low across much of the central and eastern part of the study area. The depositional environment of these fine-grained deposits is difficult to explain due to the lack of good 3D outcrop. However, the increase in relative sea level interpreted at the base of **P2** may have displaced the palaeoshoreline a considerable distance up-dip, resulting in clastic starvation of the shoreline prior to renewed progradation. The presence of a delta or series of deltas in the area during this time as outlined by Vos (1981) would provide an ample source of fine-grained detrital sediment in pro-delta or distal mouth bar settings. It was proposed by Seidl and Rohlich (1984) that the clays in the lower part of **P2** were deposited under conditions of poor circulation in a lagoon. However, Seidl and Rohlich (1984) and Parizek et al. (1984) did not describe the geographic location and distribution of the barrier islands or offshore bars which shielded these lagoonal areas. This study also failed to identify barrier islands or offshore bar deposits that may have sheltered lagoonal or swamp areas during the deposition of **P2**. The bimodal palaeocurrent data collected from some of the sandstone bodies in **P2** indicate the bidirectional movement of bars, probably during the ebb and flood of tidal currents. These bar forms recording oscillatory current action may have been of sufficient size to absorb wave energy and provide sheltered, low energy areas. However, sedimentary structures of this type were not identified during this study or by Van Houten and Karasek (1981), Parizek et al. (1984), and Seidl and Rohlich (1984). According to

Wright and Coleman (1973) and De Castro et al. (1991) a wide shelf profile similar to that interpreted from this study will have greatly reduce nearshore wave energy. It is possible that the process of wave amplification outlined by Wright and Coleman (1973) occurred in localised areas where irregularities in the shoreline profile caused wave orthogonals to converge (Jago 1980).Therefore, the physical processes normally associated with the formation of barrier islands and lagoons are probably limited in terms of their magnitude and geographic distribution, possibly restricted to areas of delta lobe abandonment. De Castro et al. (1991) state that wide shelves preclude storm activity yet favour tidal action. However, they interpret the shelf profile of the Murzuk Basin during the Middle and Upper Devonian as being narrow rather than the very wide profile interpreted in this work (Fig 1.12) and Guerrak (1991).

Within **PS1** there is no evidence that longshore currents were an important mechanism for sediment transport. Consequently, as the primary source of sediment supplied to barrier islands is from longshore currents according to Sexton and O'Hayes (1996), it is unlikely that barrier islands will have formed during the deposition of **PS1**. The low abundance and diversity ichnofabric observed in the fine-grained shallow marine deposits in the lower part of **P2** may indicate that conditions for burrowing organisms were adverse. These conditions may have been due to the physicochemical properties of the water and sediment, for example; temperature, hypersalinity, anoxia, dysaerobic conditions and high freshwater influx, or a high rate of fine-grained sediment supply. The end of **P2** is characterised by a regionally traceable surface across which there is a sharp change in facies and grain-size. This is interpreted as a parasequence boundary and represents a surface across which there is an increase in relative sea level.

Parasequence 2 (**P2**) is overlain by parasequence 3 (**P3**). In the west of the study area **P3** comprises open marine facies while marginal marine to delta plain facies were deposited in the east (Fig. 4.15). The metre thick inclined sandstone bodies observed in the upper part of **P3** at locality 21 are interpreted as prograding delta front sandstones. At locality 24 **P3** comprises shallow marine mudstones and sandstones. The coarsening-upward nature of **P3** is interpreted to be the result of the progradation of distal delta lobes or ebb tidal channels that were subsequently colonised by an ichnofabric that has obscured the primary sedimentary structures. At locality 24 **P3** can be subdivided into two coarsening-upward cycles, possibly indicating the presence of a fourth parasequence. However, this subdivision is not undertaken due to limited data on the nature and importance of the surfaces in **P3** at locality 24. The sedimentary and biogenic structures in **P3** at locality 22 indicate deposition in a minor distributary or tidal channel. The sedimentary and biogenic structures in the upper part of **P3** at locality 31 suggest deposition in a low energy tidal flat/floodplain cut by shallow sandstone-filled tidal channels. The ferruginous cemented sandstones and rootlet horizons are interpreted as exposure surfaces formed during periods of prolonged non-deposition, analogous to exposure surfaces with a similar character outlined by Bennaceff et al. (1971) and Millson et al. (1996). Palaeocurrent data from planar and trough cross-bedding, as well as from the interference and lunate ripple forms in **P3** at locality 31, give evidence of bedform migration towards the NE. These palaeocurrent data define the down dip flow direction of the meandering channels within which laterally accreting point bars migrated. The NE directed flow was probably the result of a combination of ebb tides and fluvial/deltaic currents. According to Wright and Coleman (1973) and De Castro et al. (1991) wide shelf areas are dominated by tidal processes and low wave energy, according to Elliott (1986) these conditions can produce deep, ebb dominated channels which extend seaward for considerable distances. However, there is little evidence in **P1**, **P2** or **P3** for the large channel margin bars which according to Elliott (1986) are commonly associated with these ebb channel systems. In the present day Bay of Bengal the scouring effects of strong tidal currents have resulted in the formation of broad tidal flats dissected by distributary channels which form wide, deep, straight braided estuaries (Selley 1988). According to Imperato et al. (1988) ebb and flood tidal deltas form seaward and landward respectively, of tidal inlets on microtidal and mesotidal shorelines as a result of the interaction of waves, tidal action and the effect on these factors of the inherited palaeotopography. The morphology, relative size and ratio of ebb to flood delta is controlled by the oceanographic setting of the area in which the tidal inlet was situated (Hayes 1980). The predominance of ebb tidal palaeocurrent indicators in the deltaic sandstones indicates a mesotidal range in the area (tidal range=2-4 metres) and points to the low wave energy conditions associated with such ebb tidal deltas (Hayes 1980). Palaeocurrent data from the sediments deposited within these ebb tidal inlets or deltas suggests that these morphological features are oriented perpendicular to the palaeoshoreline, a situation that also suggests deposition on a tide dominated shoreline (Caplan and Moslow 1999).

On the modern coast of South Carolina, the position and spacing of ebb tidal deltas and tidal inlets is strongly influenced by the position and geometry of inherited Holocene incised valleys (Sexton and Hayes 1996). A lowermost Middle Devonian palaeotopography has been identified by Seidl and Rohlich (1980) and this study that could have directly controlled the position of tidal inlets and associated ebb tidal deltas. The palaeotopography inherited during the lowermost Middle Devonian almost certainly caused irregularities in the palaeoshoreline which would have developed wave energy gradients, possibly forming small, sub-aqueous channel margin bars to the deltas. Although the position of ebb deltas and inlets can remain fixed, their position and lateral frequency will be strongly influenced by storm activity acting upon the marginal channel bar forms and by the lateral migration and progradation of the delta channels (Finley 1977; Hayes 1980). According to Finley (1977) the lateral migration of ebb deltas can result in the preservation of thick, laterally continuous sandstone bodies that contain bedforms with predominantly offshore (ebb) directed palaeocurrents. The palaeocurrents measured in the sandstones of **P1**, **P2**, and **P3** are generally directed offshore and may in part comprise ebb tidal delta deposits. Imperato et al. (1988) state that every 200-500 years there is a major shift in the position of the distal part of the main ebb channel due to the hydraulic inefficiency of the channel from terminal lobe deposition. Low rates of creation of accommodation space will also promote the rapid lateral incision of the channel margin bar once the available accommodation space offshore has been filled. A model for the lateral migration of these ebb tidal deltas is outlined in figure 4.16 and may be very important during the Middle Devonian due to the low rate of creation of accommodation space and low angle basin slope. At locality 36 **P3** is interpreted to comprise delta front mudstones overlain by tidal channel and mouth bar sandstones that are separated by a minor surface of incision.

During the deposition of **P1, P2,** and **P3** it is interpreted that river dominated deltas dominated palaeoshoreline processes (Fig. 4.15) with abandoned delta lobes reworked by tidal and biogenic activity. The passage from **P1,** and **P2** to **P3** in **PS1** records an overall increase in grain-size throughout the region, which reflects an increase in depositional energy and a shallowing-upwards of water depth. The coarse-grained sedimentary rocks in the upper part of **PS1** were deposited in the delta front, offshore transition zone, shoreface and foreshore in the west of the study area. While in relatively up-dip areas in the east of the study area deposition occurred in distributary channels, tidal channels and tidal flats.

The parasequences in **PS1** result from the migration of facies belts which are the result of changes in relative sea level (Fig. 4.17). The possible origins of these relative sea level fluctuations are discussed in section 4.7.

4.1.5. *Parasequence set 2 (PS2). General description.*

At locality 21 **PS2** comprises a single coarsening-upward succession of poorly exposed shales and siltstones which are overlain by thickly bedded medium to coarse-grained sandstones, forming inclined lenticular bodies that dip to the north (Fig. 4.18). The uppermost sandstones also contain numerous internal erosion surfaces and shallow channels (Fig. 4.19). The sandstones contain a low to medium diversity ichnofabric which is confined to bedset-bounding surfaces and foresets and mainly comprises *Planolites*, although rare *Tigillites* and *Thalassinoides* were also observed. The sandstones also rarely contain the external moulds of shells and the upper units of many of the sandstones contain asymmetrical ripples. The poorly exposed lower section of fine-grained sedimentary rocks and the massively bedded upper sandstones at locality 21, as illustrated in figure 4.4, make the identification of parasequences in **PS2** at this locality impossible, if they are indeed present.

At locality 24 PS2 contains parasequence 4 (P4), parasequence 5 (P5) and parasequence 6 (P6); the fine-grained elements of these cycles are generally poorly exposed. The lower parasequence (P4) terminates with asymmetrical ripples and contains ferruginous-stained shell moulds and an ichnofabric comprising *Thalassinoides, Planolites, Spirophyton* and possibly *Cruziana.* The upper two parasequences (P5 and P6) are well exposed, the lower fine-grained elements comprise purple siltstones that contain a medium abundance ichnofabric including *Planolites, Thalassinoides* and *Chondrites,* along with ferruginous-stained shell moulds, plant debris, and the imprint of a Ctenacanthus spine (Fig. 4.20). Both parasequences coarsen-upward into siltstones then sandstones which contain asymmetrical ripples while P5 comprises sandstones with low angle truncation surfaces and low angle planar laminae in P6.

At localities 22 and 23 PS2 comprises two coarsening-upward cycles that can be tentatively subdivided into parasequence 4 (P4) and parasequence 5 (P5). The parasequence boundary is placed +19 metres up the measured section where there is a decrease in grain-size across a poorly exposed section (Fig. 4.4). The lower part of P4 comprises grey shales incised into by sandstones. The lower shales contain a monotype *Tigillites* ichnofabric of low abundance. The overlying sandstones are fine to medium-grained and contain numerous rounded mudstone clasts and mudstone flakes that are concentrated on foresets and bedset bounding surfaces. The upper beds within some of the sandstones also contain asymmetrical ripples, and a *Planolites* and *Tigillites* ichnofabric of low abundance. The sandstones form a series of stacked, laterally discontinuous channels which thicken upward to form an amalgamated sandstone body. The upper 50 centimetres of P4 comprise low angle planar laminated sandstones. The lower part of P5 and the parasequence boundary surface are poorly exposed but there is a reduction in grain-size across the boundary surface (Fig. 4.4). Parasequence 5 comprises fine to medium-grained, moderately sorted sandstones that contain a *Planolites* and *Tigillites* ichnofabric of low to medium abundance. The upper surfaces of the sandstones commonly contain lunate, asymmetrical and symmetrical ripples indicating variable currents. At localities 22 and 23 P5 terminates with a series of low angle planar laminated sandstones containing low angle truncation surfaces and parting lineations.

At locality 33 P4 comprises interbedded siltstones and sandstones. The lower siltstones and sandstones are poorly exposed, containing a *Tigillites* and *Planolites* ichnofabric that show lateral variations in abundance. These are incised by a 2 metre thick channel filled with moderately to poorly sorted, fine to medium-grained, planar tabular and trough cross-bedded sandstones. These sandstones also contain clast-rich horizons, some of which may be the poorly preserved external moulds of shells. The upper surface of the sandstone-filled channel is iron-stained and pitted. The channel is overlain by 2 metres of poorly exposed fine-grained sediment that is overlain by 2 metres of well sorted and well rounded fine to medium-grained sandstone with a low abundance *Planolites* and *Tigillites* ichnofabric. The sandstones form planar tabular and trough cross-bedding with rare low angle planar laminated sandstone. This upper coarsening-upward succession is tentatively identified as P5 although limited data are available on the character and importance of the surface at the base of this coarsening-upward package.

In the east at locality 31, the lower part of PS2 is poorly exposed and inferred to be fine-grained. These units are overlain by a thick sequence of moderately to poorly sorted sandstones that contain numerous internal erosion surfaces, clast-rich channel lags, plant debris, epsilon cross-bedding, and a monotype low abundance ichnofabric of *Planolites*. These units are poorly exposed and cannot be further subdivided into parasequences, if they are indeed present (Fig. 4.4).

At locality 32 the lower part of P5 comprises sandstones incising into featureless pale yellow/green mudstones. The lower 3 metres of the sandstones are fine-grained, well sorted and contain weathered-out clasts, mudstone flakes, rounded mudstone clasts, asymmetrical ripples, herringbone cross-bedding, low angle planar laminae (bound by low angle truncation surfaces), parting lineations, flame structures and convolute laminae. These sandstones are overlain by a laterally discontinuous red siltstone horizon that is incised by a 3 metre thick succession of erosively-based, poorly to moderately sorted fine to coarse-grained sandstone. The upper sandstones are planar tabular and trough cross-bedded with frequent erosion surfaces, recumbent foresets, asymmetrical ripples and possibly herringbone cross-bedding.

Immediately adjacent to locality 32 P5 was examined at locality 30, comprising fine-grained sandstones encased within and interbedded with purple/lilac mudstones. The mudstones appear blocky and have a mottled vertical pattern that may indicate the occurrence of a poorly preserved *Skolithos/Tigillites* ichnofabric. The sandstones are well sorted and form epsilon cross-bedded sandstone bodies that are inclined in a number of directions. At locality 32 the upper part of P5 comprises fine to coarse-grained sandstones that contain weathered-out clasts, ferruginous-stained bedding surfaces and rare *Planolites* burrows. The sandstones are trough cross-bedded, erosively-based, are overlain by 2 metres of grey/pink shale with rare siltstones and very fine-grained sandstone lenses which contain a low abundance *Tigillites* and *Areniculites* ichnofabric. These sandstones are occasionally stained green, possibly by glauconite. This is overlain by a 2 metre succession of sandstones which fine-upward from medium to fine-grained sandstones, occurring in erosively-based channels and as sigmoidal sandstone bodies (Fig. 4.21). The basal surface of the channel sandstone contains linear sole marks and the sigmoidal sandstone beds contain fissile, red siltstone partings. The upper 50 centimetres comprises well sorted fine-grained sandstone with low angle planar laminae, rare planar tabular and trough cross-bedding, asymmetrical climbing ripples, imprints of plant fragments (Fig. 4.22), weathered-out clasts and an occasional *Planolites* ichnofabric. The upper surface of the sandstone is intensely pitted and iron-stained, and contains a highly abundant, subhorizontal, rootlet system (J. Howell 1998, pers comm.) (figures 4.23 and 4.24) with a radial dendritic pattern.

4.1.6. *Palaeocurrent data from PS2.*

Palaeocurrent data were collected from the logged sections and detailed study areas. These data are displayed in figure 4.25. However, the comparison of these data between outcrops is problematic due to lateral and vertical facies variations. Palaeocurrent data measurements included channel bounding surfaces, planar tabular, herringbone and trough cross-bedding, low angle planar laminae, parting lineations, low angle truncation surfaces, various ripple forms (symmetrical, asymmetrical, lunate and climbing ripples), epsilon cross-bedding and inclined bedset bounding surfaces.

The variability observed in the planar tabular-cross bedding data is not unexpected as planar tabular bedform foresets can strike, and therefore migrate, anywhere between normal and parallel to the flow direction according to Selley (1988).

4.1.7. *Palaeoenvironmental interpretation for PS2.*

Parasequences 4, 5, and 6 in **PS2** were deposited in shallow marine/ marginal marine to fluvial/deltaic conditions, an interpretation of the palaeoenvironmental evolution of the area is given in figure 4.15. The palaeocurrent indicators suggest that the regional dip was generally towards north with the palaeoshoreline oriented NNE/SSW to NNW-SSE. The palaeoenvironmental interpretation for **PS2** outlined in figure 4.15 is incorporated into figure 4.26 which displays the migration of depositional systems across the region during the deposition of **P1**, **P2** and **P3**.

At locality 21 **PS2** comprises fine-grained pro-delta sediments which are overlain by inclined sandstone bodies which are interpreted to be delta front deposits, delta top and shoreface sandstones (Fig. 4.15). The asymmetrical ripples in the large inclined sandstone bodies are interpreted to have been formed by NE and NW directed currents active in the delta front or shoreface. Alternately the ripples may represent the action of longshore currents or wind action. The restricted ichnofabric in **PS2** may be the result of adverse environmental conditions, possibly due to the introduction of large amounts of freshwater to the area by fluvial or deltaic systems (Fig. 4.15).

At locality 24 **P4** comprises a succession of mudstones and siltstones deposited in the prodelta or interdistributary bay area (Fig. 4.15). These are incised by a sandstone-filled channel complex. The sandstones within the channels contain a diverse ichnofauna including abundant *Spirophyton* burrows indicating a strong marine influence. The channel complex may represent a series of tidal inlets, ebb tidal deltas or mouth bar deposits. The channel complex is overlain by **P5** and **P6** which coarsen-upward from very fine to fine-grained sandstone and contain a diverse and abundant ichnofabric. Parasequence 6 also contains a Ctenacanthus spine indicating that the area was connected to open marine conditions as Ctenacanthus have only been found in marine or deltaic mixed strata (J. Maisey 1997, pers comm.). The presence of asymmetrical ripples migrating towards the NE indicates current activity in this direction, probably of the same origin as the structures observed at locality 21 (fluvial/deltaic, ebb tidal, longshore currents or wind action; Fig. 4.9) while the upper part of **PS2** at locality 24 comprises lower shoreface to foreshore sandstones.

At localities 22 and 23 **PS2** comprises a succession of mudstones deposited in the delta front, interdistributary bay or delta plain (Fig. 4.15). The latter palaeoenvironment may be responsible for the low abundance and diversity of the ichnofabric, although adverse palaeoenvironmental conditions could occur in any of these environments. The fine-grained units are incised by a series of E-W orientated sandstone-filled channels. These channels are interpreted as distributary channels that are initially of limited lateral extent, becoming wider and multistorey in the lower to middle part of **PS2**. The orientation of asymmetrical ripples from fine-grained sediments deposited between the channels indicates current activity towards the NNE possibly as overbank, interdistributary bay or crevasse splay deposits. In the lower to middle part of **PS2** the distributary channels are capped by a thin succession of low angle planar laminated sandstones which are interpreted to have been deposited in the lower shoreface to foreshore (Fig. 4.15). These sandstones were deposited as a result of tidal reworking after channel abandonment and may represent parasequence boundary surfaces, therefore subdividing **PS2** at localities 22 and 23 into **P4** and **P5**. The upper part of **PS2** comprises sandy tidal flat and tidal channel deposits capped by lower shoreface/ foreshore sandstones that contain symmetrical and lunate ripples indicating oscillatory flow conditions. The uppermost sandstones are also ferruginous-stained and cemented, interpreted to have formed as a result of periodic sub-aerial exposure similar to those outlined by Bennaceff et al. (1971) and Millson et al. (1996).

At locality 33 the upper part of **PS2** strongly resemble the facies observed in the upper part of **PS2** at localities 22 and 23 and are therefore placed in **P5**. These units comprise tidal flat and lower shoreface to foreshore deposits (Fig. 4.15). Palaeocurrent data (herringbone cross-bedding) indicate that tidal currents were active but bedforms were generally migrating towards the W or SW.

At locality 31 **PS2** comprises poorly exposed fine-grained sediments that are interpreted to be fine-grained delta plain deposits (Fig. 4.15). These fine-grained units are cross-cut by tidally influenced channels and associated laterally accreting point bar deposits which are erosively overlain by distributary channel sandstones. The ichnofabric in the channelled and laterally accreting sandstones indicates unfavourable conditions for colonisation, probably related to a freshwater influx to the area. The tidal channel has a coarse basal lag that contains plastically deformed grey siltstone clasts, indicating that erosional reworking of the fine-grained deposits beneath and lateral to the channels occurred.

The upper part of **PS2** at locality 32 comprises sandstones interpreted as foreshore and upper shoreface deposits (Fig. 4.15), the lateral extent of these deposits is unknown. These are overlain by lagoonal/ mudstones or tidal flats that are incised into by fluvial or delta top sandstones (Fig. 4.15). Bedforms in the fluvial and delta top sandstones are migrating towards the west indicating drainage this direction.

Immediately adjacent to locality 32 the upper part of **PS2** at localities 28 and 30 comprises sandstones and mudstones deposited in a tidal flat which is cut by tidally influenced channels that contain laterally accreting point bar bedforms (Fig. 4.15). The accretion surfaces can be split into discrete point bar forms which have variable dip directions formed during the lateral migration of numerous point bar forms in the tidal flat. The upper surface of one bar form is ferruginous-stained and pitted indicating sub-aerial exposure soon after abandonment. The upper surface of the bar form also contains a radial dendritic rootlet system (Fig. 4.23 and Fig. 4.24) which indicates the colonisation of the abandoned bar form by vegetation, possibly analogous to the in-situ *Lycophyts* (*Stigmaria*) observed by Parizek et al. (1984) and Seidl and Rohlich

(1984) in the Upper Devonian Tarut Formation approximately 20 kilometres to the east (section 4.7).

The parasequences in **PS2** (**P4**, **P5** and **P6**) record the migration of facies as a result of fluctuations in relative sea level (Fig. 4.26), the origins of which are discussed in section 4.7.

4.2. Idri Formation (Givetian to lower Frasnian).

The Idri Formation crops out along the southern margin of the Gargaf Arch immediately overlying the B'ir al Qasr Formation. The lower contact with the underlying B'ir al Qasr Formation is irregular with up to 1 metre of relief at outcrop and the upper sandstones of the underlying B'ir al Qasr Formation are intensely iron-stained, a feature that commonly indicates a period of prolonged sub-aerial exposure according to Millson et al. (1996). Previous work by Parizek et al. (1984), Seidl and Rohlich (1984) and Mergl and Massa (1992) on the southern flank of the Gargaf Arch identified an oscillation of environments and energy levels within the Idri Formation (section 4.8). The Idri Formation can be subdivided into two parasequence sets, parasequence set 3 (**PS3**) and parasequence set 4 (**PS4**) (Fig. 4.27). Parasequence set 3 contains a maximum of four parasequences (**P7**, **P8**, **P9** and **P10**) (Fig. 4.27), although it appears that the number of parasequences is modified by localised depositional controls. At locality 21 four parasequences are interpreted while at localities 17 and 18 two and three parasequences respectively are identified. It is interpreted that **P7** is localised at locality 21. However, it is recognised that the character of the lower two parasequences in **PS3** are very similar and the occurrence and preservation of **P8** may vary, rather than **P7**. Parasequence set 4 contains a maximum of three parasequences (**P11**, **P12** and **P13**) (Fig. 4.27) although at locality 17 it is possible that **P11** may be composed of more than one cycle. A graphic log correlation panel of the Idri Formation can be seen in figure 4.28, which also displays palaeocurrent data from the logged sections and detailed study areas and the position of these areas in outcrop. A colour coded basin profile has been constructed for deposition of **PS3** and **PS4** upon which the migration of facies belts have been superimposed (Fig. 4.5). A list of detailed study areas, logged sections and their locality numbers can be found in appendix 4.2. In the following text, data and their interpretations are listed to form a west to east transect of the outcrop in an up-dip and along-strike traverse.

4.2.1. *Parasequence set 3 (PS3). General description.*

At locality 41 (log 30) in the west of the study area **P7** comprises shales and siltstones overlain by low angle planar laminated sandstones that contain an abundant ichnofabric comprising *Spirophyton, Zoophycos* (Fig. 4.29), *Tigillites, Planolites* and *Thalassinoides.* The lowermost sandstones at locality 41 also contain shell moulds and plant remains.

The lower part of **PS3** was examined at locality 47 (Fig. 4.30) where **P7** and **P8** comprise two coarsening-upward shale or siltstone to sandstone cycles. The sandstones contain climbing ripples and are overlain by sandstones containing faint low angle laminae which are overprinted by a high abundance monotype *Skolithos* ichnofabric (figures 4.31 and 4.32). The upper surfaces of the sandstones capping each cycle are intensely ferruginous-stained and irregular on the centimetre scale (Fig. 4.33).

At locality 45 (log 32) **P7** and **P8** comprise heterolithic siltstones, sandstones and shales. Some of the lower sandstones contain an ichnofabric of low to medium abundance comprising *Planolites, Chondrites, Cruziana, Rhizocorraluim* (Fig. 4.34), *Nerietes (?)* and *Tigillites.* The sandstones contain mudstone clasts and mudstone flakes throughout indicating erosional reworking of fine-grained sediments, and the upper beds of many of the sandstones contain symmetrical or asymmetrical ripples.

Further to the east, at locality 42 (log 31) (Fig. 4.35) **P7** comprises poorly exposed fine-grained sediments which are incised by a thin, laterally discontinuous erosively-based sandstone-filled channel. The channel fill comprises medium to coarse-grained moderately to poorly sorted sandstones which contain numerous mudstone clasts in the lower part of the channel and mudstone drapes on many of the foresets.

At locality 43 (500 metres SE of locality 42) **P7** is well exposed and comprises mudstones, heterolithic and lenticular bedded sandstones and siltstones and erosively-based sandstone-filled channels (Fig. 4.35). The heterolithic units contain an abundant ichnofabric comprising *Rosselia* (Fig. 4.36; D. McIlroy 1997, pers comm.), *Planolites*, *Tigillites* and *Ophiomorpha* with laterally discontinuous erosively-based sandstone lenses containing the *Skolithos* ichnofabric. The lenticular-bedded sandstones contain asymmetrical ripples, hummocky cross-stratification (HCS) and numerous weathered-out clasts throughout. The lower surfaces of the sandstone-filled channels commonly contain basal groove casts and sole marks.

At locality 31 (appendix 4.1) the lower part of **P7** was observed overlying **P6** and comprises fine to very coarse-grained, poorly to moderately sorted sandstones that commonly contain a highly abundant, monotype *Zoophycos* ichnofabric. The sandstones form planar tabular and trough cross-bedded units. There is a sharp increase in grain-size and a decrease in the quality of the sorting of the sandstones across the lower contact of **PS3** with **PS2**.

At locality 41 **P7** is overlain and/or incised by a thick succession of stacked erosively-based sandstone-filled channels. The sandstones form large planar, trough and epsilon cross-bedforms with occasional clast-rich horizons. The uppermost sandstones are covered by a heavy desert varnish, therefore the identification of parasequences in this succession is problematic due to the lithology and the lack of identifiable parasequence boundary surfaces.

At locality 47 **P8** and **P9** comprise coarsening-upward siltstone to sandstone cycles. The siltstones contain a medium abundance ichnofabric comprising *Planolites*, *Tigillites* and *Diplocraterion* while the overlying sandstones contain a monotype *Skolithos* ichnofabric similar to those observed in **P7** (figures 4.31 and 4.32). The upper surfaces of the sandstones capping the coarsening-upward cycles are intensely ferruginous-stained and also contain moderately abundant examples of a *Spirophyton* ichnofabric.

At locality 45 (log 32) **P9** comprises a coarsening-upward mudstone, siltstone to fine-grained sandstone succession. The mudstones and siltstones are poorly exposed while the sandstones contain mudstone clasts, interference ripples and low angle planar laminae. This succession is overlain by **P10**, the lower part of which is poorly exposed. The upper part of **P10** comprises a thick succession of sandstones that are poorly sorted, occasionally clast-rich, ferruginous and contain numerous, large fragments from Devonian plants (up to 80 x 15 centimetres in size) (Fig. 4.37). No ichnofabric was observed in **P9** or **P10**.

At locality 42 **P8** comprises interbedded shales, siltstones and sandstones that contain an abundant ichnofabric comprising *Spirophyton*, *Chondrites* and *Diplocraterion*. This succession is overlain by a succession of massive sandstones that incise up to 1 metre into the underlying deposits, with the incision surface forming a concave-upward scour surface that can be traced for many tens of metres laterally. The sandstones are 6 to 7 metres thick and fine-upward from very coarse-grained sandstones at the base to medium-grained sandstones at the top (Fig. 4.28); no ichnofabric was observed. The basal lag of the massive sandstone is poorly sorted and very coarse-grained, containing plastically deformed clasts of the underlying shale (Fig. 4.38) which indicate that scour occurred soon after deposition of the underlying sediments. Parasequence 8 (**P8**) is overlain by **P9**, the latter parasequence comprising a thin succession of interbedded grey/green shale and fine-grained sandstones which contain *Chondrites* burrows. These fine-grained units are incised by a succession of erosively-based sandstones, the fill of which comprises 2 to 3 metres of medium to coarse-grained, planar tabular and trough cross-bedded sandstone. The sandstones also contain numerous, large flakes and clasts of mudstone (up to 80 millimetres long), the latter concentrated along foresets in the sandstones. These foresets are also commonly picked out by mudstone drapes (Fig. 4.39).

4.2.2. *Palaeocurrent data from PS3.*

Palaeocurrent data were collected from the logged sections and detailed study areas. These data are displayed in figures 4.28 and 4.40. However, the comparison of these data between outcrops is problematic due to lateral and vertical facies variations. Palaeocurrent data measurements included channel bounding surfaces, planar tabular and trough cross-bedding, low angle planar laminae, various ripple forms (symmetrical, asymmetrical, lunate and interference ripples), epsilon cross-bedding.

The variability observed in the planar tabular data is not unexpected as planar tabular bedform foresets can strike, and therefore migrate, anywhere between normal and parallel to the flow direction according to Selley (1988).

4.2.3. *Palaeoenvironmental interpretation for PS3.*

In the west of the study area **P8** and **P9** were deposited in a shallow marine to offshore shelf environment which was affected by occasional storm activity. In up-dip areas further east, one or more deltas prograded from the SE or E to the NW or W respectively (figures 4.41 and 4.42). The palaeocurrent data in **P8** and **P9** indicates the down-dip migration of bedforms towards the west with the palaeoshoreline orientated broadly N/S. In the west of the study area at locality 41 units in **P7** or **P8** are interpreted as shallow marine sandstones deposited between the offshore transition zone and lower shoreface. In the central part of the study area at locality 45 during **P7** and **P8** sedimentation occurred between lower foreshore to shallow shelf water depths, comprising erosively-based sandstone-filled channels which show a fining-upward of grain-size. These sandstone-filled channels may be the result of prograding lagoons or estuaries, tidal inlets or ebb tidal deltas. At localities 42 and 43 in the western part of the study area **P7** comprises shallow shelf sediments which are incised and overlain by erosively-based sandstone-filled channels. These are interpreted as tidal inlets or ebb tidal delta channels (Fig. 4.16), scouring into the fine-grained, shallow marine sediments beneath. The groove and flute casts on the lower surfaces of several of these channels indicates that moderate to high energy conditions prevailed during the initial incision. The symmetrical wave ripples in the upper beds of one channel fill complex indicate that deposition occurred in moderate to shallow water marine conditions subject to wave activity. The uppermost surfaces of several of the channel fill sandstones are ferruginous-stained and pitted (Fig. 4.33). As the degree of iron-staining and pitting is minor, and there is no evidence of widespread erosion, these surfaces probably formed as a result of exposure to circulating O_2 rich waters and bioturbation during a period of non-deposition. This non-deposition may have resulted from an areal shift in sandstone deposition associated with conditions of relative sea level rise, rather than the result of subaerial exposure of the sediments (section 4.7).

The upper part of **PS3** comprises shallow marine to marginal marine and non-marine facies (figure 4.41). These deposits shallow-upward from shallow shelf to delta front sediments in the west and delta front/delta top to fluvial sediments in the east. In the west of the study area at locality 41 the middle and upper part of **PS3** comprises shallow marine sandstones which pass vertically into marginal marine and delta top deposits with laterally accreting bedforms (Fig. 4.41). Palaeocurrent data from the delta top facies predominantly indicate the migration of bedforms towards the west although planar tabular data also indicate the action of east/west oscillatory tidal currents. The epsilon cross-bedding surfaces indicate the migration of laterally accreting point bars in meandering deltaic or tidal channels. The restricted ichnofabric generally observed in the uppermost sandstones indicates unfavourable conditions for colonisation, possibly due to high levels of freshwater in the area. At localities 42, 45 and 46 in the central part of the study area, **P9** and **P10** comprise shales, siltstones and sandstones deposited in a shoreface and tidal flat setting, overlain by braidplain and braid delta sediments deposited in erosively-based channels. The palaeocurrent data from within these fluvial/deltaic sediments indicates bedform migration, and probably a flow direction towards the west. These palaeocurrent data also show a limited amount of oscillatory current flow that may be due to the occasional impingement of a tidal influence near the bayline or the migration of antidunes during periods of high discharge. The abundant plant remains in **P10** at locality 45 indicate that vegetated areas were present either in swamps similar to modern mangroves, or where stable mid channel bars or overbank areas were colonised by plants.

At localities 42 and 43 in the western part of the study area **P9** and **P10** comprise a series of delta front to delta top distributary channels which incise into unconsolidated delta front and delta top sediments. The palaeoenvironment during **P9** and **P10** throughout the area is dominated by fluvial processes, in sharp contrast to the open marine conditions interpreted during the deposition of **P8**.

The parasequences in **PS3** record the migration of facies belts as a result of a fall in relative sea level, the origins of which are discussed in section 4.7 and displayed in figure 4.42.

4.2.4. *Parasequence set 4 (PS4). General description.*

The contact between **PS3** and **PS4** is represented by an abrupt change from the thick succession of sandstones in upper part of **PS3**, to shales in the lower part of **PS4**. The outcrop stratigraphy of **PS4** can be seen in figure 4.28.

At locality 41 the lower part of **P11** is poorly exposed and scree covered. Where exposed the lower part of **P11** comprises yellow/green mudstones with occasional laterally discontinuous siltstones and very fine to fine-grained sandstones. Several of the sandstone beds contain HCS (Fig. 4.43), and the lower surfaces of the sandstones contain prod marks and sole marks oriented N-S and NW-SE. The sandstone are erosively-based and are laterally discontinuous across 100 metres of outcrop. Due to poor exposure the ichnofabric could not be examined in detail but comprises low abundance *Planolites* and *Tigillites* where the interbedded shales and sandstones are exposed. The middle part of **P11** comprises grey/green micaceous siltstones which are incised by thin, laterally discontinuous low angle planar laminated medium-grained sandstones, the upper surfaces of which may be HCS. The sandstones contain a restricted ichnofabric of occasional *Planolites* and *Tigillites*; rare plant material is also present. Laterally the sandstones grade into siltstones which contain an increased abundance of the ichnofabric seen in the sandstones. The upper part of **P11** comprises siltstones and sandstones. The siltstones are micaceous and contain an abundant ichnofabric comprising *Planolites*, *Tigillites* and *Thalassinoides*. The siltstones are incised by a channel with up to 1 metre of erosional relief at its base. The channel fill comprises fine to medium sandstone containing: a) rare external moulds of large shells, b) a diverse and abundant ichnofabric (as in the underlying siltstones), c) a basal lag containing rounded clasts of the underlying siltstone, and d) internal bedding surfaces containing parting lineations oriented N/S.

At locality 42 **P11** comprises poorly exposed grey/green shale which immediately overlies the ferruginous-stained, irregular upper surface of **PS3** (Fig. 4.28). These lower shale units are incised by a 2 metre thick medium-grained sandstone, the lower beds of which contain numerous rip-up clasts of the underlying shale. These clast-rich sandstones are planar and trough cross-bedded and contain mudstone-filled scours on the foresets. The upper sandstone beds contain symmetrical and asymmetrical ripples with mudstone lenses between the ripple sets. The sandstones also contain *Planolites* burrows.

At locality 41 **P11** is overlain by **P12**, the latter comprising shales overlain by very fine to medium-grained sandstones. While the shales are poorly exposed, the sandstones contain an intense ichnofabric comprising *Planolites*, *Tigillites*, *Skolithos* and *Thalassinoides*. The uppermost sandstone contains a highly abundant *Skolithos* ichnofabric, the upper surface of which is iron-stained.

At locality 47 the lower part of **PS4** was examined and is believed to be comparable to **P11** or **P12**. This succession comprises interbedded shales, siltstones and sandstones incised by a 19 metre thick sandstone succession. The shale, siltstone and sandstone interbeds contain a highly diverse ichnofabric comprising *Planolites*, *Chondrites*, *Tigillites*, *Thalassinoides* and *Arenicultes*. The sandstones are fine to coarse-grained and contain low angle planar laminae and rare planar and trough-cross bedding. The sandstone also contain an ichnofabric of low to high abundance comprising *Planolites* and *Skolithos*.

At locality 42 **P12** comprises 3 metres of poorly exposed white/grey shale that coarsens-upward into siltstones which are interbedded with sharp based very fine to fine-grained sandstones. The fine-grained sandstones contain planar and trough-cross bedforms, ferriginous nodules, and a low to very high abundance, low to medium diversity ichnofabric of *Planolites*, *Tigillites* and *Skolithos*. The sandstones are thinly bedded, contain asymmetrical ripples, and in places form laterally discontinuous sandstone-filled channels, one of which contains a highly abundant monotype ichnofabric of *Skolithos*.

At locality 46 a succession comprising thinly bedded well sorted fine-grained sandstones with tuning fork and straight crested symmetrical and asymmetrical ripples. The lower sandstones form large planar tabular and trough-cross bedforms containing mudstone clasts and a low abundance ichnofabric comprising *Planolites*, *Tigillites* and *Areniculites* (Fig. 4.44), separated by low angle bedset bounding surfaces. This succession may correspond to **P12**.

At locality 41 the lower part of **P13** is poorly exposed but is overlain by interbedded sandstones and shales, terminating in a 1 metre thick, erosively-based channel filled by medium-grained sandstone. The internal character of the upper sandstone is unknown as desert varnish has obscured any sedimentary or biogenic structures that may have been present.

At locality 44 the upper few metres of **PS4** are exposed and comprise poorly sorted, fine to coarse-grained sandstones that contain numerous erosion surfaces and associated coarse to granular lags. The sandstones are planar and trough cross-bedded which contain a low abundance *Planolites* and *Tigillites* ichnofabric. The uppermost 1 metre of sandstones at locality 44 contains an intense ferruginous cement and is stained dark brown. These sandstones were probably deposited during the upper part of **P13**.

At locality 42 **P13** comprises interbedded sandstones and siltstones incised by a 10 metre thick sandstone succession (Fig. 4.28). The interbedded sandstones and siltstones contain a medium to very high abundance ichnofabric comprising *Planolites*, *Tigillites*, *Chondrites*, *Diplocraterion* and *Skolithos*. The sandstones in the channel are fine to coarse-grained and are planar and trough cross-bedded. The sandstones also contain numerous scour surfaces, asymmetrical ripples, lateral accretion surfaces and laterally discontinuous, poorly sorted lag deposits on scour surfaces.

At locality 46 the upper metre of **PS4** comprises occasional siltstones and very fine-grained sandstones. These uppermost sandstones contain low angle planar laminae, tuning fork ripples (Fig. 4.45) and a low to very high abundance ichnofabric, comprising *Planolites*, *Tigillites*, *Areniculites*, *Chondrites* and *Thalassinoides*. This succession is interpreted to be comparable to **P13**.

4.2.5. *Palaeocurrent data from PS4.*

Palaeocurrent data were collected from the logged sections and also from detailed study areas, these data are displayed in figures 4.28 and 4.46. The comparison of these data between outcrops is problematic due lateral and vertical facies variations. Palaeocurrent data measurements included planar tabular and trough-cross bedding, groove casts, sole marks, low angle planar laminae, parting lineations, low angle truncation surfaces, symmetrical and asymmetrical ripples, epsilon cross-bedding, and undifferentiated cross-bedding. Hummocky cross-stratification (HCS) were also observed although these data were not collected for analysis as they are of limited use for interpreting palaeocurrents.

The variability observed in the planar tabular data is not unexpected as planar tabular bedform foresets can strike, and therefore migrate, anywhere between normal and parallel to the flow direction according to Selley (1988).

4.2.6. *Palaeoenvironmental interpretation for PS4.*

Parasequence set 4 (**PS4**) conformably overlies **PS3** in the study area with no evidence of major erosion between the two. The parasequences in **PS4** in the central and western part of the region were deposited in a shallow marine environment (Fig. 4.41) with the NNE to SSW oriented palaeoshoreline situated up-dip to the SE.

At locality 41 **P11** and **P12** are interpreted to have been deposited in moderate water depth marine conditions. The HCS sandstones in **P11** indicate deposition above the storm wave base but the palaeoenvironment of the lower part of **PS4** cannot be further constrained. In the west of the study area **P13** may have been deposited in shallow marine conditions. However, the palaeoenvironment of these units is poorly constrained due to poorly exposed section. It is interpreted that parasequence set 4 records a shallowing of water depth at locality 41.

In the east of the study area at locality 42 **P11** and **P12** were deposited in a shallow to marginal marine setting and are interpreted as marginal marine tidal flat and delta front/top deposits. The poorly exposed fine-grained units may represent tidal flat (mudstone dominated), interdistributary bay or delta plain deposits (Fig. 4.41). The sandstones are interpreted to have been deposited within minor channels seaward of the bayline, possibly in and around the shoreface in tidal channels as indicated by the *Skolithos* ichnofabric (Fig. 4.29). The mudstone lenses between ripple sets and mudstone-filled trough scours are interpreted to have been deposited as the depositional fluctuated during tidal activity.

The uppermost parasequence in PS4 is **P13**, comprising shallow marine and fluvial/deltaic deposits in the west and east of the study area respectively (Fig. 4.41). In the west of the study area at localities 41 and 47 **P13** is interpreted to have been deposited in shallow marine conditions. At locality 41 the bioturbated sandstones and siltstones are interpreted as offshore transition zone and lower shoreface deposits. The iron-stained upper surfaces of the sandstones in **P13** were probably the result of prolonged contact with oxygen rich waters during periods of non-deposition and contact with oxygenated waters. The uppermost sandstones of **P13** at locality 41 are interpreted to have been deposited in a minor channel formed in the lower shoreface, incising into fine-grained offshore transition zone deposits. This channel may represent a tidal inlet with an associated ebb tidal delta and tidal channel (Fig. 4.16). At locality 47 **P12** comprises foreshore and shoreface deposits. Also in the west of the study area the upper few metres of **P13** at locality 44 were probably deposited in a braidplain or delta system. The sandstones do not contain any sedimentary or biogenic structures to directly indicate any marine influence but probably correlate with the shallow marine deposits observed in the uppermost part of **PS4** at locality 41. In the west of the study area at locality 42 the upper part of **PS4** comprises shoreface deposits overlain by braided and meandering fluvial/deltaic sandstones (Fig. 4.41).

In the eastern part of the study area the upper part of **PS4** at locality 46 comprises sandstones deposited in a channel draining to the SW. The channel is interpreted as a delta top distributary channel with associated interdistributary bay, delta plain and delta top sediments present laterally (Fig. 4.41). The channel is overlain by sandstones that contain symmetrical and interference ripples, and a diverse and abundant ichnofabric which are interpreted as tidal flat and shoreface deposits (Fig. 4.41). These sediments are interpreted to have been deposited during a phase of channel abandonment and were reworked by marine processes. The succession at locality 46 is interpreted to be comparable to **P12** and **P13** at other localities.

The parasequences in **PS4** record the migration of facies belts as a result of fluctuations in relative sea, these are displayed in figure 4.47.

4.3. Quttah Formation (Frasnian).

The Quttah Formation crops out along the southern margin of the Gargaf Arch immediately overlying the Idri Formation. The lower contact with the underlying Idri Formation is irregular with several metres of erosional relief measured at outcrop by Seidl and Rohlich (1984) and Parizek et al. (1984). The Quttah Formation comprises a single parasequence set (**PS5**) (Fig. 4.48), recognised as an asymmetrical (negative) cycle by Seidl and Rohlich (1984). Within **PS5** a maximum of five parasequences can be identified (**P14, P15, P16, P17** and **P18**) (Fig. 4.48) although their correlation is problematic due to the limited dataset and sandstone dominated succession in the central and eastern part of the study area. In the east of the study area at locality 25 the middle and upper part of **PS5** comprises a succession of sandstones. The subdivision of these sandstones into parasequences is problematic. However, some of the iron-cemented horizons are believed to represent parasequence boundary surfaces and can be used to identify **P15, P16, P17** and **P18**. A graphic log correlation of **PS5** can be seen in figure 4.49, which also displays palaeocurrent data from the logged sections and detailed study areas. This figure also illustrates the position of these locations. A colour coded basin profile has been constructed for the deposition of **PS5** upon which the migration of facies belts during the deposition the Quttah Formation have been superimposed (Fig. 4.5). A list of detailed study areas, logged sections and their relative locality numbers can be seen in appendix 4.3. In the following text, data and their interpretations are listed in a west to east traverse that cuts across the outcrop in an up dip and along strike sense.

4.3.1. *Parasequence set 5 (PS5). General description.*

At locality 48 (log 42) in the west of the study area **P14** and **P15** comprise shale to sandstone coarsening-upward successions, with the fine-grained sediments poorly exposed (Fig. 4.49). The sandstone beds are usually sharp based and contain low angle planar laminae, rare symmetrical ripples and variable amounts of articulated and fragmented brachiopods and bivalves. The upper few centimetres of the sandstone beds are hummocky cross-stratified (HCS), the lower surfaces commonly display sole marks and groove marks and a grain lineation can be observed along cleavage planes. The siltstones and sandstones commonly contain a low to high abundance ichnofabric comprising *Planolites*, *Tigillites*, *Chondrites* and *Bifungites Fezzanitus*.

In the east of the study area at locality 46 (log 33) the lower part of **P14** is poorly exposed. In the same area at locality 49 (log 41) **P14** comprises interbedded shales and siltstones with rare sharp based sandstones in the upper few metres. A laterally discontinuous, poorly sorted coarse-grained basal sandstone was also observed at locality 49, directly infilling the irregular upper surface of **P13** (Fig. 4.49). The shales and siltstones in **P14** at locality 49 contain a low to medium abundance ichnofabric comprising *Planolites* and *Skolithos* (D. McIlroy 1998, pers. comm) (Fig. 4.50) and possibly *Chondrites*, *Ophiomorpha* and *Areniculites* while the upper surfaces of the siltstones are pitted and undulose. The sandstones are fine-grained, sharp based, and contain sole marks groove marks on their lower surfaces and rare examples of the *Bifungites Fezzanitus* ichnofabric. At localities 22 and 25 the upper part of **P14** comprises erosively-based medium to coarse-grained sandstones. These sandstones are planar tabular and trough cross-bedded and contain numerous rounded and angular mudstone clasts. The uppermost sandstones of **P14** at localities 46 and 49 contain large ferruginous nodules and iron-cement (Fig. 4.51). This succession is overlain by **P15** which comprises sandstones with subordinate siltstones. The sandstones contain low angle planar laminae, planar tabular and trough cross-bedding, and the upper beds contain an intense iron-cement and ferruginous nodules similar to those outlined in figure 4.51.

In the west of the study area at locality 48 **P16** comprises a coarsening-upward shale, siltstone, to sandstone succession. The lower units are poorly exposed, comprising grey/green shale which do not contain clear sedimentary or biogenic structures, these shales are overlain by interbedded siltstones and sandstones. The siltstones contain an abundant ichnofabric comprising *Planolites*, *Tigillites*, *Chondrites* and *Thalassinoides*. The majority of the sandstones contain low angle planar laminae and abundant articulated and fragmented brachiopods and bivalves that can be seen in figure 4.52. A small fragment of *Agnathan* dermal armour was also discovered (Fig. 4.53) (identified by J. Maisey 1997). The sandstones also contain rare interference, tuning fork, and asymmetrical ripples in the upper metre or so. These sandstones are capped by an erosively-based sandstone-filled channel. The basal surface of the channel has up to 1 metre of incision on it. However, laterally the amount of incision decreases to almost zero. Where the incision occurs the channel is filled by ferruginous, coarse-grained, poorly sorted sandstone, that contains mudstone clasts and abundant fragmented and articulated brachiopods and bivalves. Laterally and immediately overlying the erosive channel, planar tabular cross-bedded sandstones occur, containing rare mudstone clasts and shell moulds. Lateral to the channel axis (area of maximum incision) the mean grain-size decreases from upper medium to very fine-grained sandstone.

In the east of the study area at locality 49 **P16** comprises fine to coarse-grained sandstones which are planar tabular and trough cross-bedded and contain symmetrical rippls. No ichnofabric was observed in **P16**.

At locality 48 in the west of the study area **P17** and **P18** comprise shale to sandstone coarsening-upward successions. The shales are commonly poorly exposed and do not contain clear sedimentary or biogenic structures. The grain-size of the sandstones increases towards the top of the parasequences, as does the bed thickness. The sandstones in **P18** contain symmetrical ripples, low angle laminae, low angle truncation surfaces, articulated shells, shell debris and a low to very high abundance and medium to very high diversity ichnofabric comprising *Planolites*, *Tigillites*, *Chondrites*, *Thalassinoides*, *Diplocraterion* and *Areniculites*. The sandstones in **P18** comprise a 2 metre thick succession of planar tabular and trough cross-bedded sandstones that are poorly exposed due to desert varnish.

At locality 49 **P17** and **P18** comprises a succession of planar tabular and trough cross-bedded sandstones which contain subordinate low angle planar laminae, epsilon cross-bedding and symmetrical ripples. The upper few metres of **P17** at locality 49 also contain rare examples of a low abundance ichnofabric comprising *Planolites* and *Tigillites*, and abundant external moulds of brachiopods and bivalves (Fig. 4.54).

An incomplete section of the upper part of **PS5** was also observed at locality 50 where intensely bioturbated very fine-grained sandstones are interbedded with poorly sorted, fine to coarse-grained sandstones and well sorted, low angle planar laminated fine-grained sandstones. This succession is probably equivalent to **P18** at locality 49.

4.3.2. *Palaeocurrent data from PS5.*

Palaeocurrent data were collected from the logged sections and detailed study areas. These data are displayed in figures 4.49 and 4.55. However, the comparison of these data between outcrops is problematic due lateral and vertical facies variations.

Palaeocurrent data measurements include planar tabular and trough-cross bedding, groove casts, sole marks, low angle planar laminae, parting lineations, low angle truncation surfaces and symmetrical, asymmetrical and interference ripples. Hummocky cross-stratification (HCS) were also observed although these data were not collected for analysis as they are of limited use for interpreting palaeocurrents.

The variability observed in the planar tabular data is not unexpected as planar tabular bedform foresets can strike, and therefore migrate, anywhere between normal and parallel to the flow direction according to Selley (1988).

4.3.3. *Palaeoenvironmental interpretation for PS5.*

In the east and west of the study area **P14** and **P15** at localities 46, 48 and 49 comprises open marine facies deposited in shallow shelf water depths above the storm wave base (Fig. 4.56). The sandstones are interpreted to be the result of storm activity which has re-mobilised sandstones deposited in shallow water causing the sandstones to scour into the fine-grained deeper water deposits. The abundant articulated and fragmented brachiopod and bivalve and brachiopod shells are interpreted to have been broken up and transported from shallow water shell beds into deeper water by the storms. The upper units of these sandstones were deposited in the waning stages of these storms which were reworked by late stage storm currents to produce HCS. Unusually these storm deposits are not capped by wave ripples (J. Howell 1999, pers comm.), indicating their deposition occurred below fairweather wave base or in shallower water depths where the effects of wave action were minimal. The pitted upper surfaces of many of the siltstone beds are interpreted to be the result of a prolonged period of reworking by burrowing organisms during low rates of sedimentation. This bioturbation occurred after the introduction of the coarser sediment to the sea floor by storm activity (Fig. 4.56). The coarsening-upward parasequences (**P14** and **P15**) may have formed by the progradation of the distal part of a delta lobe, ebb tidal delta, or shoreface/foreshore complex. The orientation and position of the of the palaeoshoreline during deposition of **PS5** is difficult to ascertain due to the limited dataset but is probably oriented NE-SW as interpreted from the underlying parasequence sets.

In the west of the study area at locality 48 **P16** comprises open marine facies deposited in shallow shelf water depths above the storm wave base, possibly above the fairweather wave base in the upper few metres of the succession. The abundant articulated shells, fragmented shells and the fragment of *Agnathan* dermal armour probably originated in shallower water shell banks, reworked by storm currents (Fig, 4.53). Meanwhile in the east of the study area at locality 49 facies in **P16** are interpreted as fluvial or delta top channel sandstones.

In the west of the study area at locality 48 **P17** is interpreted to have been deposited in fluctuating energy conditions in and around the lower shoreface and offshore transition zone. These units comprise bioturbated tidal flat, shoreface and foreshore deposits containing a number of low angle truncation surfaces. The uppermost sharp based sandstone in **P18** may represent the sandy fill of a tidal channel or tidal delta. However, these sandstones are poorly exposed.

In the east of the study area at localities 46 and 49 **P17** and **P18** comprise thick sandstones that generally contain little in the way of palaeoenvironmental indicators. However, the symmetrical ripples and the thin lense of *Skolithos* borrowed sandstone in **P17** at locality 46 indicate a marine influence. Palaeocurrent data demonstrates the migration of planar tabular and trough bedforms as well as rare laterally accreting bar forms. The lack of definitive data limits the interpretation to a high energy palaeoenvironment, possibly in a braidplain or braid delta with occasional meandering channel point bar sandstones. Sedimentological and faunal data from **P17** and **P18** at locality 49 are interpreted to indicate deposition in shallow to marginal marine conditions (Fig. 4.56). Sedimentary structures and palaeocurrent data indicate that shoreface sandstones were deposited in areas adjacent to active deltaic complexes that were migrating towards the NW. These data suggest that the upper part of **PS5** comprises marine influenced delta top, tidal creek and possibly shoreface deposits (Fig. 4.56).

The lower, middle and upper parts of **PS5** record the migration of facies belts as a result of fluctuations in relative sea level, these are displayed in figure 4.57.

4.4. Dabdab Formation (Frasnian).

The Dabdab Formation crops out along the southern flank of the Gargaf Arch immediately overlying the Quttah Formation. The lower and upper contacts were not examined due to the generally poor exposure of the Dabdab Formation, time limitations and problems accessing certain areas of outcrop (section 2.1). A graphic log of the measured section of the Dabdab Formation can be seen in figure 4.58, which also illustrates the position of the logged section in outcrop. The log of the

Dabdab Formation illustrated in figure 4.58 comprises data from a number of outcrops and is unlikely to be a complete measured section through the Dabdab Formation. Seidl and Rohlich (1984) observed a 12.5 metre measured section of the Dabdab Formation while Van Houten and Karasek (1981) recorded approximately 20 metres of section through equivalent age sediments. It appears likely from the work of Van Houten and Karasek (1981) that the upper part of the Formation was not observed during the course of this study at locality 51. However, the measured section does provide key data that have been used to help interpret the palaeoenvironment of the area during this time. The measured section of the Dabdab Formation comprises two parasequences, these are parasequence 19 (**P19**) and parasequence 20 (**P20**) and are placed in parasequence set 6 (**PS6**). A list of detailed study areas and their locality numbers can be seen in appendix 4.3.

4.4.1. *Parasequence set 6 (PS6). General description.*

At locality 51 **P19** comprises several metres of apparently featureless pale brown/grey mudstones that are overlain by 3 metres of interbedded, ferruginous cemented, very fine to fine-grained sandstones, siltstones, ferruginous oolitic sandstones and dark brown clay/mudstone (Fig. 4.59). A lateral transition from ferruginous cemented sandstones into ferrugiinous oolitic sandstone is commonly observed with the ferruginous oolitic sandstones occurring in lenses. The ferruginous sandstones and oolites contain a low to medium abundance ichnofabric comprising *Tigillites, Planolites, Thalassinoides, Skolithos* and possibly *Phycodes.* The siltstones are pale cream in colour and contain a moderately abundant ichnofabric comprising *Tigillites, Thalassinoides* and *Chondrites* while the mudstones appear to be barren of any ichnofabric. Sedimentary structures were generally not observed in the ferruginous sandstones or oolitic sandstones, possibly overprinted by the ichnofabric or not studied due to the limited accessibility of lateral outcrops. However, rare poorly preserved low angle planar laminae were observed in the very fine to fine-grained ferruginous sandstones. The uppermost surface of **P19** is heavily pitted, irregular and ferruginous-stained.

At locality 27 **P20** comprises 2.8 metres of apparently featureless grey mudstone which is very poorly exposed (Fig. 4.58). This is overlain by 1.2 metres of ferruginous oolite with occasional interbeds of grey mudstone. The oolite comprises ooids and peloids which are equivalent in grain-size to very fine or fine-grained sandstone, these occur in a ferruginous siltstone matrix. Lateral variations in oolite and peloid concentrations are common, a decrease from 25 % of the bulk rock total to <5 % in 30 metres of lateral exposure was observed. The ferruginous oolites contain a monotype *Skolithos* ichnofabric, which varies from low to high abundance across 10 metres of lateral exposure. Sedimentary structures were not observed in the ferruginous oolites, possibly overprinted by the ichnofabric or not studied due to the limited study of lateral outcrops.

4.4.2. *Palaeocurrent data from PS6.*

Palaeocurrent data from **P19** and **P20** at locality 51 was limited to the rare poorly preserved low angle laminae from the ferruginous-stained very fine to fine-grained sandstones. These sedimentary structures can be deposited during high flow conditions. However, the paucity of these and other palaeocurrent data make refined interpretation impossible.

4.4.3. *Palaeoenvironmental interpretation for PS6.*

The two parasequences in **PS6** at locality 51 comprise bioturbated lenticular beds of ferruginous oolite and ferruginous sandstones that are interbedded with fine-grained sediments. These data are interpreted to suggest that these facies were deposited in a shallow marine environment where the rate of siliciclastic sediment supplied to the area was moderate to low (Fig. 4.60). A specific palaeoenvironment of deposition for the ferruginous oolites in **PS6** is difficult to establish. However, a synopsis of ferruginous oolite formation is given in section 3.7.2. The sedimentological and faunal data from the logged section at locality 51, and previous studies by Bellini and Massa (1980), Van Houten and Karasek (1981), Seidl and Rohlich (1984), Parizek et al. (1984), Guerrak (1991), and Pierobon (1991), interpret that the sedimentary rocks within **PS6** were deposited in a shallow marine setting, proximal to a relatively low energy coastline (Fig. 4.60). This coastline received abundant detrital iron and large quantities of fine-grained sediment according to Van Houten and Karasek (1981). Mergl and Massa (1992) concur that the faunal content of the Dabdab Formation suggests a lagoonal to semi-lagoonal palaeoenvironment, with a restricted connection to the open sea. The vertical repetition of the laterally discontinuous ferruginous oolite shoals and siliceous sandstones, that are underlain by mudstones and clays at locality 51, indicate that both oolitic and siliciclastic sediment supply varied through time. These variations in grain-size and facies may have resulted from relative sea level fluctuations, the origins of which are discussed in section 4.7. A number of coarsening-upward cycles are recognised in coeval sedimentary rocks, interpreted as the product of the repeated avulsion of a major river away from this area of the palaeocoastline (Van Houten and Karasek (1981).

The lack of palaeocurrent data and limited overall study of **PS6** during this study have made the identification of the location and orientation of the palaeoshoreline problematic. Previous work by Van Houten and Karasek (1981) interpreted a NE source of sediment during deposition of the Dabdab Formation. This source of sediment would result in a palaeoshoreline oriented broadly NW-SE with drainage systems perpendicular to this source area. The interpretations of Van Houten and Karasek (1981) are in contrast to the E or SE source of sediment that is interpreted in this study for the underlying parasequence sets (**PS1** to **PS5**) of Mid to Late Devonian age (sections 4.1 to 4.5).

The mean grain-size of the sediment and the amount of coarse-grained siliciclastic material within **PS6** is greatly reduced when compared to the underlying Middle and Upper Devonian deposits. The change in the physiochemical composition of the sediment deposited is interpreted to be the result of a major change in the depositional processes acting in this area of North Africa, the possible causes of which are discussed in section 4.9 and chapter 5. The lack of data from lateral sections of the Dabdab Formation mean that the detailed interpretation of relative sea level changes within the formation cannot easily be made.

4.5. Tarut Formation (Fammenian).

The Tarut Formation crops out along the southern flank of the Gargaf Arch immediately overlying the Dabdab Formation. The lower contact of the Tarut Formation with the underlying Dabdab Formation was not studied due to time limitations and restrictions in accessing certain areas of outcrop (section 2.1).

A graphic log of the measured section of the Tarut Formation measured at locality 52 can be seen in figure 4.58 which also displays the location the study area. The measured section of the Tarut Formation comprises a single parasequence set containing two parasequences, parasequence 21 (**P21**) and parasequence 22 (**P22**) (Fig. 4.58). The boundary between the two parasequences in **PS7** that comprise the Tarut Formation is difficult to identify. However, for the purpose of this study the parasequence boundary is placed on top of the last sandstone in the lower part of the succession, approximately 9.5 metres from the base of the logged section at locality 52 (Fig. 4.58). A list of detailed study areas and their locality numbers can be seen in appendix 4.3.

4.5.1. *Parasequence set 7 (PS7). General description.*

At locality 52 **P21** comprises a succession of 10 metres of mudstones which are interbedded with sandstones and siltstones, the lower 3 metres of which are poorly exposed (Fig. 4.61). The mudstones are varicoloured and do not contain any sedimentary structures or ichnofabric. The siltstones are varicoloured, micaceous and commonly contain fine to medium grains of sandstone. The sandstones comprise very fine to fine-grained quartz sandstone which contain a low abundance ichnofabric comprising *Skolithos Rugosus*, *Treptichnus* (Fig. 4.62) (D. McIlroy 1998, pers comm.) and *Planolites*. The sandstones also contain complete and fragmented shell material which are generally confined to the lower few centimetres of the sandstone beds as lag deposits. The sandstones occur in sharp based sheet and laterally discontinuous bodies (Fig. 4.63) which fine laterally to siltstones; these incise into the underlying mudstones. The sandstones also contain low angle planar laminae, hummocky cross-stratification and asymmetrical ripples. The lower surfaces of the sandstones also contain groove marks, prod marks, and evidence of basal scour.

At locality 52 **P22** comprises a succession of interbedded clays, siltstones, sandstones, oolites and micro-conglomerates (Fig. 4.58). The lower few metres comprise claystones and purple siltstones which contain a monotype *Tigillites* ichnofabric. These units are overlain by varicoloured siltstones that contain a medium to very high abundance ichnofabric that is dominated by *Skolithos* traces (Fig. 4.64) although *Planolites*, *Tigillites*, *Chondrites*, *Thalassinoides*, *Phycodes*, and *Diplocraterion* are also observed. The siltstones contain occasional interbeds of fine-grained sandstone which contain faint low angle planar laminae and are re-worked by the abundant ichnofabric in the surrounding siltstones. The upper 2 metres of **PS7** contains erosively-based oolitic sandstone bodies that pass laterally and vertically into conglomerates (Fig. 4.65). The oolitic sandstones are red/purple in colour and contain rounded mudstone clasts, rare bone debris, plant material and a monotype *Skolithos* ichnofabric. The oolitic sandstones are rarely planar tabular and trough cross-bedded. In one of the oolitic sandstone beds the *Skolithos* traces have been deformed and are deflected 15° from vertical (Fig. 4.66). The conglomeratic units comprise a fine-grained oolitic sandstone matrix with intraclasts of sandy oolite, ferruginous-stained brachiopods, bone fragments, plant material and rounded mudstone clasts (Fig. 4.65). The clasts in the conglomerate are up to 30 centimetres in diameter. The conglomerates commonly contain a low abundance monotype *Skolithos* ichnofabric.

4.5.2. *Palaeocurrent data from PS7.*

Palaeocurrent data from **P21** were restricted to rare groove marks, and asymmetrical ripples, while palaeocurrent data from **P22** comprises planar tabular and trough cross-bedding. Hummocky cross-stratification was also observed in **P21**. The dataset is limited due to the facies present and small data set, therefore the results are inconclusive.

4.5.3. *Palaeoenvironmental interpretation for PS7.*

The palaeoenvironmental evolution of **PS7** can be split into two, with the lower and upper parasequence deposited in distinctly different palaeoenvironments. The location and orientation of the palaeoshoreline during deposition of **PS7** is difficult to ascertain from the limited data. Previous work by Van Houten and Karasek (1981), Parizek et al. (1984), and Seidl and Rohlich (1984) suggests a NE source for sediment during this time with the palaeoshoreline oriented broadly NW-SE.

The lower parasequence (**P21**) at locality 52 (log 61) is interpreted to have been deposited in open marine conditions above the storm wave base (Fig. 4.67). The fine-grained, background sedimentation is interbedded with sandstones that contain sedimentary structures and biogenic material (HCS, basal scour, groove marks, fossiliferous basal lags) which are interpreted to be the result of storm activity (Fig. 4.67). The palaeocurrent data from this part of the succession is sparse, indicating only that NW/SE directed currents scoured into the substrate. As these are interpreted as storm generated currents, and associated deposits, these data are of little use in determining the local basin dip and drainage direction, probably only indicating a component of the storm and geostrophic currents active during this time. The storm origin of many of these lower sandstones is also recognised by Van Houten and Karasek (1981), De Castro et al. (1991), and Pierobon (1991), but not by Parizek et al. (1984), or Seidl and Rohlich (1984).

At locality 52 parasequence 22 (**P22**) is interpreted to have been deposited in shallow marine conditions in and around the foreshore and shoreface that were receiving limited amounts of siliciclastic sediment (Fig. 4.67). The generally abundant and diverse ichnofabric indicates there was an abundant supply of nutrients to the area. The *Skolithos Rugosus* bioturbated oolitic sandstones and siltstones are interpreted to represent fine-grained bars deposited in shallow water depths. It is unusual for this ichnofabric to be found in fine-grained sediment (D. McIlroy 1998, pers comm.) and may represent a number of firmground surfaces formed during periods when sediment supply to the region was greatly reduced. The presence of the deformed *Skolithos* burrows (Fig 4.66) indicates that the upper surface of the sediment was sometimes subject to shear stresses, possibly analogous to the recumbent foresets commonly observed in fluvial and shallow marine siliciclastic deposits such as those displayed in figure 3.11. The conglomerates that occur laterally and immediately overlying the oolitic sandstones are interpreted to be tidal delta, delta front or delta top distributary channel deposits. The intraclasts of oolitic sandstone in the conglomerate indicate erosional re-working of the underlying oolitic sandstones, while the presence of abundant bone fragments indicate the presence of vertebrates in up-dip areas, or the re-working of bone debris in older deposits. The preservation state of the bone fragments indicates some degree of transport and fragmentation, possibly due to transport into this area from an up-dip or lateral source. The large size of the intraclasts in the conglomerate indicate that the depositional energy was high. These deposits also contain rare *Skolithos* burrows which may still indicate a marine influence. However, the depositional energy was high enough to restrict the colonisation of the sediment by a diverse ichnofauna. The faunal content of the upper part of the Tarut Formation was also interpreted by Mergl and Massa (1992) to indicate a mixed, shore-related palaeoenvironment. The plant material in the oolitic sandstones and conglomeratic layers indicates that area's up-dip or laterally were colonised by plants. Examples of in-situ *Lycophyts* (*Stigmaria*) are recorded on the southern flank of the Gargaf Arch by Sedil and Rohlich (1984) and Parizek et al. (1984). Van Houten and Karasek (1981) also report macerated plant debris in the claystones that dominate the middle part of the Tarut Formation at locality 52.

The logged section of PS7 at locality 52 is interpreted as a shallowing-upward succession, a process which is illustrated in figure 4.68.

4.6. Ashkidah Formation (Fammenian to Tournaisian).

The Ashkidah Formation crops out along the southern flank of the Gargaf Arch immediately overlying the Tarut Formation (PS7). The lower contact and lower 3 metres of the Ashkidah Formation were studied at locality 52 and can be seen in figure 4.58, which also illustrates the location of this study area. The main part of the Ashkidah Formation at locality 52 and laterally equivalent sections were not studied due to time limitations. However, previous work by Van Houten and Karasek (1981), Parizek et al. (1984), and Seidl and Rohlich (1984) have provided further data on this level of the stratigraphy. The lithological composition of the Ashkidah Formation outlined below, and its relationships with the underlying PS7, has lead to the Ashkidah Formation being placed in a new parasequence set 8 (PS8). A list of detailed study areas and their locality numbers can be seen in appendix 4.3.

4.6.1. *Parasequence set 8 (PS8). General description (lowermost 3 metres only).*

The lower 3 metres of PS8 at locality 52 comprises interbedded, lenticular, and flaser bedded sandstones, siltstones, and mudstones (Fig. 4.69). The lower contact with PS7 is sharp and irregular on the decimetre scale, representing an abrupt change from the conglomeratic oolites in the upper part of PS7 (Fig. 4.65), to the fine-grained heterolithic sandstones, siltstones, and mudstones in the lower part of PS8 (4.69). The upper surfaces of many of the sandstones are symmetrically ripped, low angle planar laminated sandstones are also observed. The sandstones and siltstones contain a low to high abundance and low to medium diversity ichnofabric that mainly restricted to the upper surfaces of the sandstones. The ichnofabric comprises *Planolites*, *Tigillites*, *Diplocraterion*, and *Areniculites*.

4.6.2. *Palaeocurrent data from PS8.*

Palaeocurrent data from PS8 were limited to the symmetrical ripples and low angle laminae in the sandstones. The former dataset indicates oscillatory flow conditions, while the latter sedimentary structures can form in upper or lower plane bed flow regime (Leeder 1983). The limited palaeocurrent dataset has limited further interpretation of paleoflow conditions and the overall palaeoenvironment during PS8.

4.6.3. *Palaeoenvironmental interpretation for PS8.*

Although the data collected from PS8 at locality 52 is limited to the lower few metres of the sequence/parasequence set it can be used in conjunction with previously published data to interpret the palaeoenvironment during the deposition of PS8. The lower units of PS8 at locality 52 comprise heterolithic units which contain a moderately diverse and abundant ichnofabric and evidence of oscillatory flow conditions. As such the lower units of PS8 are interpreted to have been deposited in a shallow to moderate water depth marine environment, probably above the fairweather wave base (Fig. 4.70). The marine palaeoenvironment appear to have received moderate amounts of fine-grained siliciclastic sediment in contrast to the predominantly fine-grained and oolitic sediments observed in PS7. The open marine facies in the lower part of PS8 overlie the marginal marine to continental conglomerates in the upper part of PS7 and thus indicate a period of relative sea level rise across the PS7 and PS8 boundary. The abrupt change in lithofacies type from PS7 to PS8 is potentially the result of increased amounts of siliciclastic sediment supplied to the area by fluvial/deltaic discharge or longshore drift mechanisms. However, an increase in relative sea level occurred across the boundary between PS7 and PS8, juxtaposing marginal marine and continental deposits with deeper water facies. Blanpied and Rubino (1998) state that the unconformable contact between PS7 and PS8 is enhanced by tectonism.

Further data is needed to accurately constrain the palaeoenvironment in the lower part of PS8 and how this evolves through the middle to upper part of PS8. Previous work by Van Houten and Karasek (1981), Parizek et al. (1984), Seidl and Rohlich (1984) and Mergl and Massa (1992) interpret oscillations in relative sea level during this time leading to offshore shelf to coastal plain conditions. Parizek et al. (1984) and Seidl and Rohlich (1984) also note an increase in continental conditions towards the east of the study area and interpret a NE source for sediment during this time (sections 4.6.3 and 4.7.3).

4.7. Sequence Stratigraphic interpretation.

The palaeoenvironmental evolution of the Middle to Upper Devonian facies on the southern flank of the Gargaf Arch can be interpreted using the sequence stratigraphic concepts and terminology outlined in section 2.4. The previously described and interpreted Middle to Upper Devonian succession (sections 4.1-4.6) has been summarised to construct a vertical profile (Fig. 4.71) within which six sequences (S8 to S13) were identified, corresponding to the six sequences of Van Houten and Karasek (1981). Sequences one to seven (S1 to S7) are contained in the Lower Devonian succession detailed in chapter 3. Within S8 to S13 eight parasequence sets (PS1 to PS8) were identified, with S8 and S9 containing two parasequence sets (PS1 to PS4) while sequences ten to thirteen inclusive contain only one parasequence set each (PS5 to PS8). The lower seven of these parasequence sets were studied in detail and are detailed in sections 4.1 to 4.6 and are illustrated in figure 4.71. The parasequence sets also contain a variable number of parasequences and are separated by surfaces that are interpreted to have a sequence stratigraphic significance. These surfaces include sequence boundaries, parasequence set boundaries and flooding surfaces (Fig. 4.71). The distinction between the sequence boundary surfaces and parasequence set boundary surfaces within the Middle to Upper Devonian succession is commonly subtle. However, the sequence boundary surfaces generally have a greater topographic relief and are very heavily pitted and iron-stained when compared to other surfaces in the succession. These sequence boundary surfaces also represent surfaces across which the interpreted change in facies is much greater than that interpreted across the parasequence set bounding surfaces in the sequences. Both the sequences boundaries and parasequence set bounding surfaces record formed as a result of changes in relative sea level, the possible driving mechanisms of which are discussed in section 4.7.3.

4.7.1. *Recognition of sequences and sequence boundaries.*

The six sequence boundaries identified in the Middle to Upper Devonian succession form the bounding surfaces of sequences eight to thirteen (S8 to S13). These sequences are shown in figure 4.71 which also illustrates the relative positions of the sequence boundaries in the Middle to Upper Devonian succession. The position of the sequence boundaries generally corresponds to the Formation boundary surfaces picked by Seidl and Rohlich (1984). The sequence boundaries were identified on the basis that they represented regionally extensive surfaces across which there was a major change in facies and commonly evidence of erosion. The erosion associated with these surfaces may be the result of subaerial processes or transgressive erosion associated with the subsequent relative sea level rise.

The lowermost bounding surface of the Middle to Upper Devonian can be traced across the entire outcrop study region. This surface separates Cambrian and Ordovician sandstones from Middle Devonian sedimentary rocks in the east and central part of the study region and Silurian shales from coeval deposits in the west of the study area. This surface is interpreted to be a sequence boundary of tectono-eustatic origin (section 1.2.3) and represents a prolonged period of erosion and non-deposition. To accommodate basin wide correlation this surface is also interpreted as the boundary between the Lower and Middle to Upper Devonian succession in other areas of the basin although this relationship is poorly understood. This is therefore labelled as sequence boundary seven (**SB7**), with its origin and nature further discussed in chapters 5 and 6.

Sequence boundaries eight, nine, ten, eleven and twelve (**SB8**, **SB9**, **SB10**, **SB11** and **SB12**) can be traced throughout the study region on the southern flank of the Gargaf Arch. The possible origins of these surfaces are discussed in chapter 5. There remains a degree of uncertainty on the placement of several of the sequence boundary surfaces at a number of localities. Sequence boundary eleven was not directly studied due to poor exposure. However, it is identified as a sequence boundary for the reasons outlined above.

The variations in accommodation space associated with relative sea level fluctuations during the Middle to Upper Devonian resulted in the formation of sequences that contain sediments which can be placed within Transgressive, Highstand, Falling Stage and possibly Lowstand Systems Tracts (TST, HST, FSST and LST respectively). The lack of abundant LST deposits in this region throughout the Devonian is discussed in sections 2.4.6 and 3.9.7, with the possible driving mechanisms of these relative sea level fluctuations outlined in section 4.7.3.

4.7.2. *Parasequence sets and parasequences.*

Within the eight parasequence sets outlined above a number of parasequences are also identified. The number and character of these parasequences in many of the sequences and parasequence sets varies in a lateral and down-dip sense, the possible origins of which are discussed below. The juxtaposition and character of these parasequences and parasequence sets can be used to interpret the systems tracts within which they occur, therefore enabling accommodation space variations to be studied.

<u>Parasequence set 1 (PS1): (Lower part of Sequence eight).</u>

At the base of **PS1** in the west of the study area a basal lag has been identified which is absent from the central and western parts of the study area. This basal lag is interpreted to have been deposited during the early stages of a period of relative sea level rise during the early part of **PS1**. The basal lag in **PS1** is therefore interpreted to have been deposited late during a LST or TST. This rise in relative sea level led to the transgression of a pre-Middle Devonian palaeotopography that developed during the formation of sequence boundary seven **SB7**. This lag is of restricted areal distribution, and was deposited in localised pockets of increased accommodation space or areas of preferential preservation. The localised areas of increased accommodation space may represent topographic lows in a pre-Middle Devonian palaeotopography generated during the formation of **SB7**. The decrease in absolute accommodation space towards the east may also be due to the influence of structural features or variations in the basement rheology (section 5.7).

The main part of **PS1** comprising parasequences 1 to 3 (**P1**, **P2** and **P3**) is 18 metres thick in the west and at least 23 metres thick in the east. Localised thickness variations occur with **PS1** 12 metres thick at locality 24, which is only 20 kilometres east of an 18 metre thick succession of **PS1** at locality 21. Parasequences 1 to 3 generally have well developed coarsening-upward signatures and are between 2.5 and 9 metres thick. Parasequence 1 generally terminates with a surface across which there is an abrupt decrease in grain-size throughout the region, this is especially marked in the east of the study area (figures 4.4 and 4.71). This may represent the maximum flooding surface in **PS1** which is overlain by **P2** and **P3**. The overall character of **P2** and **P3** appears to be aggradational in the west and progradational in the east, therefore these parasequence are interpreted to have been deposited during a HST (Fig. 4.71). The facies juxtaposition during **PS1** can be seen in figure 4.17. Parasequence set 1 is capped by a regionally traceable surface across which there is an abrupt decrease in grain-size and a major decrease in siliciclastic sediment supply.

Parasequence set 2 (PS2): (Upper part of Sequence eight).

The lower bounding surface of **PS2** comprises a regionally traceable surface across which an increase in palaeowater depth is interpreted. This is interpreted to be a flooding surface that formed as a result of a rise in relative sea level (Fig. 4.71). This transgressive event caused a shut-off of siliciclastic sediment supply to the study area and shifted the palaeoshoreline up-dip to the SE. Within **PS2** two or three coarsening-upward parasequences are identified (**P4**, **P5** and **P6**). At locality 21 parasequences in the delta front sandstones that make up the upper part of **PS2** are difficult to identify but interpretation of photo montages of comparable levels of the succession laterally suggests that a number of parasequences are present (figure 4.2b). The stacking pattern of the parasequences in **PS2** appears to be progradational, deposited when the rate of relative sea level rise was low or during a stillstand. Parasequences 4, 5 and 6 are therefore interpreted to have been deposited during a HST, although the poorly exposed lower part of **PS2** may contain a fine-grained succession deposited during a TST. The identification and correlation of parasequences in **PS2** is complicated by the presence of the erosively-based sandstone-filled distributary channels in the central and eastern part of the study area during the deposition of **PS2** (figures 4.4, 4.71 and 4.72). These distributary channels had the potential to locally erode out a number of parasequences as they prograded basinward during conditions of low relative sea level rise, stillstand or more likely, relative sea level fall (Fig. 4.73). Using the definitions outlined in section 2.4.2 and appendix 2.1, the erosively-based distributary channels in **PS2** are interpreted to have formed during a late HST or during the FSST of Plint

(1997a; 1997b). The FSST distributary channels may also be responsible for the absence of pro-delta and delta front sediments in the central and eastern part of the study area, eroding the latter facies during progradation (Fig. 4.73). There is no evidence that these distributary channel sandstones are LST incised valley fill deposits, on the contrary, the plastically deformed clasts that commonly line the base of these channel sandstones are composed of the same material that underlies the channels (figures 4.4 and 4.38). The composition and preservation state of these clasts suggests that the period of incision occurred almost contemporaneously with the deposition of the fine-grained, sub-channel sandstone succession. PS2 is capped by a regionally traceable surface which has several metres of erosional relief and across which there is an abrupt decrease in coarse-grained sediment supply to the area. This is interpreted as sequence boundary eight (SB8) (Fig. 4.71).

Parasequence set 3 (PS3): (Lower part of Sequence nine).

The lower bounding surface of PS3 is sequence boundary eight (SB8) which, as well as being a sequence boundary, also comprises a regionally traceable surface across which there is an increase in marine influence, making it a combined sequence boundary and flooding surface. This flooding surface formed as a result of a rise in relative sea level that can be recognised throughout the outcrop study region (Fig. 4.71), the possible origins of which are discussed in section 4.7. The absence of LST deposits above the sequence boundary is discussed in section 4.7.3. The irregular nature of the upper surface of SB8 may be the result of subaerial erosion during the formation of the sequence boundary or transgressive erosion during the subsequent rise in relative sea level.

In the central and western part of the study area four parasequences are interpreted in PS3, these are parasequences 7, 8, 9 and 10 (P7, P8, P9 and P10) (Fig. 4.71), while in the eastern part of the study area three parasequences are interpreted in PS3, comprising parasequences 8, 9 and 10. The lateral variations in the number of parasequences in PS3 may be due to localised depositional effects such as erosion by the distributary channels that dominate the succession in the east of the study area (figures 4.4, 4.72 and 4.73). The sediments filling these distributary channels generally have a similar composition and character to those observed in **PS2** and are therefore interpreted as HST and FSST deposits (Fig. 4.73), incising into the finer-grained HST and TST deposits beneath. At locality 21 the lower units of the 7 metre thick sandstone present in **P11** are not observed, therefore its relationship with the underlying sediments can only be estimated. The most likely explanation is that this sandstone can be correlated with the other HST and FSST channels found laterally. However, there remains a small possibility that part of this sandstone succession represents an incised valley fill complex deposited during the succeeding LST. The parasequences in **PS3** form a progradational stacking pattern that records a fall in relative sea level and basinward shift in facies, and as such are interpreted as prograding HST and FSST parasequences (Fig. 4.73).

The upper surface of **PS3** at locality 18 is immediately underlain by 60 centimetres of sandstone that contains an intense ferruginous cement and large ferruginous nodules that are interpreted to have formed during a prolonged period when iron rich pore waters circulated within the sandstones. The upper surface of **PS3** is a regionally traceable surface across which there is an abrupt decrease in grain-size.

<u>Parasequence set 4 (PS4): (Upper part of Sequence nine).</u>

The lower bounding surface of **PS4** comprises a regionally traceable surface across which a major increase in marine influence and palaeo-water depth are interpreted (Fig. 4.71). Therefore, the lowermost units of **PS4** were deposited during a TST although they are poorly studied due to their limited exposure. A major increase in relative sea level in units equivalent to those at the base of **PS4** in the same area is also recognised by Mergl and Massa (1992) and J. Rubino (1998, pers comm.), the origins and implications of which are discussed in section 4.7.3.

In the west of the study area six coarsening-upward parasequences are identified, while a maximum of five are interpreted further to the east at locality 42 (Fig. 4.71). These parasequences form a progradational parasequence set (**PS4**) which indicates that during their deposition the rate of relative sea level rise was decreasing, at a stillstand, or falling (Fig. 4.73), making these HST, FSST or LST deposits. Bearing in mind the highly transgressive units at the base of PS4 it is more likely that deposition took place during the former two systems tracts rather than the latter. At locality 42 the upper part of PS4 comprises a 10 metre thick sandstone-filled distributary channel which is absent 1.5 kilometres to the east (up-dip). Comparable facies occur 10 kilometres to the west (down-dip). The lateral extent of these distributary channels is unknown as a result of the limited exposure of strike oriented sections. As noted in PS2 and PS3, these distributary channels have probably caused the erosion of any delta front or shallow marine parasequences deposited prior to the westward progradation and incision of the distributary channels. The lower channel contacts and internal character of the sandstone are very similar to those observed within the HST and FSST sandstones in PS2 and PS3. The erosive nature of the distributary channel sandstones in PS4 suggests that that they were deposited when accommodation space was limited and are interpreted to have been deposited late during a HST or FSST (Fig. 4.73).

The upper bounding surface of PS4 comprises a regionally traceable surface across which there is an abrupt decrease in grain-size interpreted to be the result of a major decrease in siliciclastic sediment supply to the area. This surface also has several metres of erosional relief upon it and is interpreted as sequence boundary nine (SB9) (Fig. 4.71).

Parasequence set 5 (PS5): (Sequence ten).

The lower bounding surface of PS5 is SB9, comprising a regionally traceable surface above which there is an increase in marine influence and also an interpreted increase in palaeo-water depth (Fig. 4.71). Sequence boundary nine is irregular in the central and eastern part of the study area, which may be the result of subaerial or transgressive erosion. The irregular nature of this contact throughout the region was also recognised by Seidl and Rohlich (1984) who noted the regionally disconformable nature of the contact. The flooding event that occurs in the lower part of PS5 separates distributary sandstones in the upper part of PS4 from open marine shales and hummocky cross-stratified sandstones (HCS) in the lower part of **PS5**. In the lower part of **PS5** it is difficult to pick clearly defined parasequences as the sandstones are generally of storm origin. However, five coarsening-upward cycles are recognised at locality 48 and are interpreted as parasequences in **PS5**. Due to the limited study of **PS5** the lateral correlation and comparison of these parasequences has not been undertaken. At localities 46 and 47 the open marine shales and HCS sandstones are incised by a multistorey sandstone body. At a number of levels in this sandstone body there are heavily ferruginous cemented sandstones which contain large ferruginous nodules. These iron-rich horizons may correlate with the parasequence bounding surfaces interpreted from locality 48, formed when ferruginous cement and nodules precipitated during the prolonged circulation pore waters rich in iron. The continued evidence of marine influence in the multistorey sandstone body suggests that even though relative sea level was fluctuating during the generation of parasequences, the mean relative sea level only varied slightly. This minor variation in base level is also observed at locality 48 where parasequences in the middle and upper part of **PS5** record only a minor shallowing of relative sea level, possibly during the HST and/or early FSST. The upper bounding surface of **PS5** comprises a regionally traceable surface across which there is a major decrease in mean grain-size and supply of siliciclastic sediment to the entire study area.

Due to data limitations the degree of confidence associated with the sequence stratigraphic interpretations of **PS6**, **PS7** and **PS8** is lower than the interpretations made from **PS1** to **PS5**. However, the sequence stratigraphic interpretations for **PS6** to **PS8** may still be valid and provide a framework for future study.

Parasequence set 6 (PS6): (Sequence eleven).

The lower bounding surface of **PS6** is sequence boundary ten (**SB10**), a surface across which there is a major change in depositional style throughout the region (figures 4.71 and 4.74). The change from dominantly siliciclastic sedimentation in the upper to part of **PS5** to the deposition of fine-grained sediment and ferruginous oolites in **PS6** requires a major change in sediment supply and resultant depositional processes in the marginal to shallow marine environments. Palaeoshorelines starved of siliciclastic sediment can be typical of detached Lowstand shorelines. However, for the shallow marine units of **PS6** to have been deposited above the marginal marine late HST and FSST deposits in the upper part **PS5** the intervening TST, HST, and FSST need to be absent. In this respect **PS6** was deposited under similar base level conditions as the Lower Devonian Ouan Kasa Formation (FA7) that is detailed in section 3.9.6. The starvation of siliciclastic sediment to the palaeoshoreline is interpreted to be the result of source area inactivity or a change in the character of sediment supplied to these areas. An indication of this may be the change in sediment supply direction noted between **PS1** to **PS5** and **PS6** (from the E/SE and NE respectively). The origin of the mechanism(s) driving these changes in sediment supply characteristics is difficult to ascertain without data from these areas. However, the shallow marine interpretation for **PS6** suggests that a rise in relative sea level may have been responsible for the shut off of siliciclastic sediment to the region (Fig. 4.74).

Sequence boundary ten also appears to be responsible for erosion of the underlying units in the central and eastern part of the Gargaf Arch outcrop. Parasequence set 5 thins towards this region and is completely absent from the eastern end of the Gargaf Arch (Fig. 4.49). The possible origins of this erosion, rise in relative sea level and change in the character of sediment supplied to the basin are further discussed in section 5.5.2.

The magnitude of relative sea level rise associated with the flooding surface at the base of **PS6** is difficult to ascertain as the amount of fall associated with the formation of **SB10** cannot be estimated from the data available. However, the major change in depositional style across **SB10** suggests that whatever the factors responsible for its formation, or the magnitude of relative sea level fall associated with its formation, the subsequent period of relative sea level rise resulted in the deposition of very different facies. Therefore, as the sedimentary rocks in **PS6** are interpreted to have been deposited during conditions of high or rising base level they are placed within a TST, corresponding to the interpretations of Blanpied and Rubino (1998).

In **PS6** two coarsening-upward parasequences, parasequence 19 (**P19**) and parasequence 20 (**P20**) are interpreted, the origins of which (autocyclic vs allocyclic) cannot be ascertained due to a lack of data from lateral sections. The limited data obtained from the Dabdab Formation during this study are recognised to limit accurate interpretations. However, previous work by Seidl and Rohlich (1984) interpret the Dabdab Formation as a separate Formation, and therefore for the purposes of this study the Dabdab Formation is interpreted as a separate sequence (**S11**) and parasequence set (**PS6**). The reduced thickness of **PS6** may be due to the characteristics of the relative sea level fluctuations responsible for its formation. This relative sea level fluctuation may be of a higher frequency and lower magnitude than those responsible for the formation of the over- and underlying parasequence sets and sequences. Data limitations prevent more accurate interpretations in this respect. The upper bounding surface of **PS6** was not directly observed due to poorly exposed sections and time limits on field work. Although the character of this surface is poorly constrained, previous work by Seidl and Rohlich (1984) and Parizek et al. (1984) interprets this succession as a separate cycle, and therefore a separate sequence (**PS6**).

<u>Parasequence set 7 (PS7): (Sequence twelve).</u>

The open marine shales and fine sandstones in the lower part of **PS7** were deposited above the shallow marine ferruginous oolites that make up **PS6**. The transition between these two parasequences is interpreted to represent a major rise in relative sea level from the shallow marine facies in the upper part of **PS6** to the open marine facies in the lower part of **PS7** (Fig. 4.71). As well as marking a flooding surface that can be traced across the entire region, this surface is also sequence boundary eleven (**SB 11**).

The lower units of **PS7** comprise parasequence 21 (**P21**) that was deposited when the rate of relative sea level rise was high or rising during a TST. The middle to upper parts of **PS7** record an overall shallowing of relative sea level from the open marine shales and sandstones in **P21** to the fine-grained shoreface and foreshore deposits, oolite shoals and tidal channel deposits capped by conglomerates in **P22**. Previous work by Van Houten and Karasek (1981), Seidl and Rohlich (1984) and Parizek et al. (1984) also recognised the overall shallowing-upward nature of these facies. Parasequence 22 may therefore have been deposited as the rate of relative sea level rise decreased, during a HST. The possible origins of the change in sediment supply direction noted between **PS1** to **PS5** and **PS7** (from the E/SE and NE respectively) are discussed in chapter 5.

The starvation of siliciclastic sediment from marginal and shallow marine areas during the deposition of the upper part of **PS7** is very similar to the situation interpreted to have been active during deposition of the underlying **PS6**. Therefore deposition of **PS7** took place during a period of quiescence when sediment supply to the area was greatly reduced as local sediment source areas were inactive.

<u>Parasequence set 8 (PS8): (Sequence thirteen).</u>

The lower units of **PS8** overlying **PS7** were observed at locality 52, comprising delta front/foreshore fine-grained sandstones, siltstones, and shales. Where the contact between **PS7** and **PS8** was observed at locality 52 the contact is irregular and represents a surface across which there is an abrupt increase in palaeowater depth and change in facies. This surface is interpreted as sequence boundary twelve **(SB12)** (Fig. 4.71). The rise in relative sea level associated with the deposition of **PS8** has reactivated siliciclastic sediment supply to the region, albeit in the form of fine-grained sandstones. Previous work by Seidl and Rohlich (1984), Parizek et al. (1984), and Van Houten and Karasek (1981) and more recently by Blanpied and Rubino (1998) suggest that **PS8** records a shallowing of relative sea level upwards, although there is an increase in the areas connectivity to open marine (siliciclastic) conditions to the N and E according to Seidl and Rohlich (1984).

4.7.3. *Interpreting the origin and distribution of relative sea level fluctuations in the Middle to Upper Devonian succession on the northern margin of the Murzuq Basin.*

The fluctuations in relative sea level described in subsections 4.7.1 and 4.7.2 in the Middle to Upper Devonian facies (**PS1 to PS8**; Fig. 4.71) can be interpreted in terms of their possible origins and duration. Biostratigraphic data from previous work in the study area by Seidl and Rohlich 1984, Parizek et al. 1984, Massa (1988), and Mergl and Massa (1992) have provided stratigraphic age limits for the sequences to stage level (Fig. 4.75).

The six sequences identified in the Middle to Upper Devonian succession on the northern margin of the Murzuk Basin were deposited over a maximum of 23.5 Myr (Fig 4.75); which would give an average of 3.9 Myr per sequence. However, the six sequences do not have the same sedimentological character or stratigraphic thickness and therefore it would be an oversimplification to assume their deposition occurred at a constant rate throughout the Mid to Late Devonian. The lower three sequences comprise siliciclastic deposits while the upper three sequences are predominantly fine-grained, containing claystones, siltstones, and ferruginous oolites. It is thus more reasonable to assume that the duration and rate of deposition of each sequence varied. It is also assumed the sequence boundary surfaces make up a large proportion of the duration of each sequence, possibly up to 50 % of the time available, which may be reasonable in this slowly subsiding ramp type basin (Fig. 1.12).

The sequence boundaries interpreted in the Middle to Upper Devonian succession (Fig 4.71) have a number of origins in terms of the mechanisms driving the relative sea level fall during their formation. The lowermost sequence boundary (**SB7**) appears to be a regionally important surface representing a prolonged period of relative sea level fall that was synchronous or preceded by a phase of tectonism (section 1.2.3).

The input of tectonism on the formation of **SB10** was recognised from this study, while Blanpied and Rubino (1998) interpret that **SB12** was also enhanced by tectonism. The character of the former sequence boundary differs in the western and east/central parts of the study area. In the west of the study area **SB12** represents an unconformity surface between the siliciclastic and fine-grained oolitic sedimentary rocks in **S10** and **S11** respectively. This is in contrast to the east/central part of the study area where the thickness of **S10** beneath the sequence boundary decreases rapidly and pinches out. Sequence boundaries ten and twelve may be similar to **SB7** in the respect that they formed during a fall in relative sea level that was locally enhanced by tectonism. Of the remaining sequence boundaries identified (**SB8**, **SB9** and **SB11**) **SB11** appears to be result from a period of relative sea level fall of lesser magnitude and duration, separating two sequences with a very similar sedimentological character (**S11** and **S12**) (Fig. 4.71).

According to Emery and Myers (1996) the average duration of the Middle to Upper Devonian sequences (3.9 Myr) would suggest that the duration of these sequences is controlled by either short 2nd order, or more likely, 3rd order eustatic cycles (Fig. 2.6). However, active deposition did not occur for the entire duration of each cycle. On the contrary, the sequence, parasequence set and parasequence boundary surfaces also make up an indeterminable amount of the time within each cycle. It has been established that in slowly subsiding basins the duration of the FSST and LST, and therefore the period of erosion, is increased when compared to a basin subsiding at a greater rate (Wehr 1993).

The parasequences and parasequence sets in the Middle to Upper Devonian succession indicate that higher frequency, lower magnitude relative sea level fluctuations were also active during this time. These higher frequency sea level cycles may have been of 4th, 5th or 6th order duration, acting to enhance or obscure the effects of the 3rd order cycles that were responsible for the formation of the sequences. The upper bounding surface of **S11** (**SB11**) appears to be of lesser sequence stratigraphic significance than the other sequence bounding surfaces, and may represent one such higher frequency surface 4th order cycle bounding surface.

Owing to difficulties in establishing palaeowater depth it is difficult to estimate the magnitude of relative sea level rise associated with the 3rd and 4th order cycle flooding surfaces (i.e.: overlying the sequence boundaries and parasequence sets). However, the basinward dislocation of facies (and therefore relative sea level rise) associated with the boundary between **PS3** and **PS4**, and also between **PS5** and **PS6** are interpreted to be of greater magnitude than the other flooding surfaces, possibly due to a composite 3rd and 4th order base level fall followed by a rise. The latter flooding surface represents the transition from dominantly siliciclastic (**PS1** to **PS5**) to fine-grained and ferruginous oolitic deposition (**PS6** to **PS8**). However, the base level rise associated with the end of **PS6** is interrupted by a lower magnitude base level fall which produces the parasequence set (and sequence) bounding surface between **PS6** and **PS7**. A composite sea level curve can also be seen in figure 4.76 which also outlines a possible model for the timing and duration of depositional sequences during the Middle to Upper Devonian and lowermost Carboniferous.

The driving mechanisms responsible for the formation of these relative sea level fluctuations during the Mid to Late Devonian need to be considered. As detailed in section 2.4 relative sea level fluctuations can cause the formation of depositional sequences bound by unconformities and their correlative conformities. These relative sea level fluctuations can result from autocyclic processes although the lateral tracability and cyclic nature of the Mid to Late Devonian sequences suggests other mechanisms may be responsible for their formation, namely eustasy, climate and tectonism. These factors are often interdependent (Shanley and McCabe 1994), the driving mechanisms of which are outlined in figure 2.6. According to Cloetingh (1988); Emery and Myers (1996) and Miall (1996, 1997) these 3rd, 4th, 5th and 6th order cycles result from glacioeustasy, climate or tectonism. It is difficult to envisage that the affects of episodes of tectonism with multiple orders of frequency and amplitude are recorded on this intracratonic margin of Gondwana during the Mid to Late Devonian. However, the pre-Devonian tectonism outlined in sections 1.2 to 1.3, and the influence of tectonism on the character of **SB7**, **SB10** and **SB12** in the Middle to Upper Devonian succession is poorly constrained.

4.7.4. *Regionally specific factors influencing relative sea level.*

During the Mid to Late Devonian the Murzuk Basin and surrounding region were situated on the northern margin of the Gondwanan supercontinent. Within this intracratonic margin the Murzuq Basin floor and surrounding region dipped at a low angle towards the NW and generally subsided at a low rate (section 1.4.1). The low angle of basin floor dip and generally low rate of regional subsidence affected the character of sedimentation resulting from relative sea level fluctuations in a manner unique to the Murzuq and surrounding basins in Gondwana during this time. Localised variations in the above mentioned factors have also been interpreted as a result of this study.

Lateral Variations in subsidence

The logged sections of **PS1** to **PS8** and previous work by Parizek et al. (1984) and Seidl and Rohlich (1984) have identified a regional decrease in the thickness of each sequence or formation towards the east. If the initial sequence overlying the pre-Middle Devonian palaeotopography (**PS1**, lower B'ir al Qasr Formation) was the only part of the stratigraphy to demonstrate this decrease in thickness towards the east, the control would most likely be localised erosively generated variations in accommodation space. These localised accommodation space variations could have formed during the pre-Middle Devonian tectonic activity outlined in section 1.2.3 and discussed in section 5.7. However, the decrease in stratigraphic thickness noted by previous workers including Parizek et al. (1984) and Seidl and Rohlich (1984), demonstrates that localised variations in accommodation space occurred during the majority of the Mid to Late Devonian. It is interpreted from this study that these thickness variations result from a localised area of reduced subsidence, centred in the area around Brak on the southern flank of the Gargaf Arch (Fig 4.77). There may also be evidence that the Middle to Upper Devonian sequences begin to thicken in the far west of the study area (Seidl and Rohlich, 1984), (Fig 4.78).

The predominance of facies interpreted to have been deposited during the HST and FSST are due to the reasons already outlined in sections 2.4.2 and 3.9.7.

The interpreted variations in accommodation space may have resulted in the re-orientation of the Mid to Late Devonian palaeoshoreline and drainage patterns (Fig. 4.77). These variations in the rate of accommodations space creation may also have led to the concentration of HST and FSST distributary channel systems (figures 2.12, 2.13 and 4.73) in these areas of low(er) subsidence. Palaeocurrent data from facies deposited during the FSST in the areas where the rate of accommodation space varies from the "regional" rate, may therefore not be representative of palaeocurrents active in the basin as a whole during the other systems tracts.

Chapter 5. Interpretation of subsurface data relating to the Devonian succession in the Murzuq Basin and its correlation with outcrop data.

5.1. The subsurface dataset.

A variety of subsurface data from the central area of the Murzuq Basin are presented in this chapter. These data are interpreted and correlated with outcrop data from the northern and western margins of the basin given in chapters 3 and 4, and with published data from other areas of the basin. The subsurface information mainly comprises a variety of wireline logs, seismic reflection profiles, petrographical and palynological data from sidewall core, as well as data and interpretations derived from oil and service company reports. The type and location of the wireline log data obtained from the basin are shown in figure 5.1 while the distribution and coverage of the seismic data can be seen in figure 5.2.

5.2. Interpretation of wireline log data from the central area of the Murzuq Basin.

A variety of log data from 39 wells in the Murzuq Basin were used to interpret the subsurface distribution and expression of the Devonian succession. The wireline data were obtained using several logging tools that were run in wells drilled from the 1950's to the 1990's and are therefore of variable quality and interpretative use (section 2.3.1). Data from gamma ray wireline log traces from 21 wells were used to interpret lithological variations and effect correlation of the Devonian succession in the subsurface. The gamma ray wireline data are derived from concessions NC58, NC101, NC115, NC174 and the immediately surrounding areas (Fig. 5.1). The older and poorer quality log data (section 2.3.1) are generally from the NW of the basin and therefore interpretations derived from this area may be less reliable (Fig. 5.1). The interpretations derived from these wireline data were cross checked with data from sidewall cores, core reports, well formation tops and cuttings.

5.2.1. *Gamma ray wireline data.*

Gamma ray wireline data from 21 wells in the Murzuq Basin have been correlated and used to construct four correlation panels in a variety of orientations across the basin. These panels provide a framework within which thickness variations can be recognised. Analysis of the gamma ray traces have identified two sequences in the Middle to Upper Devonian succession; these are Sequence A (SA) and Sequence B (SB) (Fig. 5.3), which can be correlated throughout the region (Fig. 5.4). In SA and SB higher resolution subdivisions can be identified utilising the shape of the wireline traces. These subdivisions include the identification of sequence stratigraphic units including systems tracts, and several key surfaces associated with them (section 2.4.4 and appendix 2.1). The distribution and internal character of the systems tracts in SA and SB are interpreted to vary throughout the region as a result of localised changes in depositional and preservational processes. Parasequences are also inferred to occur in both of the sequences. However, due to the wide spacing of the wells and the limited sidewall core data available from these wells, no attempt is made to correlate the individual parasequences.

5.2.2. *Spectral Gamma ray data.*

Spectral gamma ray data were available from well A1-NC174 and were studied in an attempt to constrain the position of key surfaces in the Devonian succession not picked by the total count gamma ray data. The spectral data from well A1-NC174 are analysed using the methods outlined by Davies and Elliott (1996), a summary of which are given in section 2.3.1.

The gamma ray wireline trace for well A1-NC174 can be seen in figure 5.4. This figure also shows the position of several sequence stratigraphic key surfaces in the succession, picked from total count gamma data. When the spectral gamma data for well A1-NC174 are examined, in conjunction with the total count data, similar interpretations can be made (Fig. 5.5). The positions of the key surfaces in well A1-NC174 remain unchanged when spectral gamma data are interpreted. However, the presence and position of a maximum flooding surface in SB is confirmed at a depth of

1641 metres where a high total count gamma peak occurs in conjunction with very high uranium readings (10 parts per million (ppm)) (Fig. 5.5). These spectral gamma ray data can be used to distinguish this as a maximum flooding surface, rather than a lesser flooding surface (Davies and Elliott 1996). The sequence boundary separating SA and SB does not have the characteristic spectral gamma response which is commonly associated with such surfaces (Davies and Elliott 1996). However, the spectral gamma response can be masked by the character of the underlying facies or the overlying TST or HST incised valley fill, TST or HST deposits (Davies and Elliott 1996). The lack of spectral gamma data from other wells mean that the character of the key surfaces in well A1-NC174 cannot be correlated.

5.2.3. *Non-Gamma ray well data.*

The data from the 18 wells that do not include gamma ray wireline data provide total thickness information for the Devonian succession as well as petrological, stratigraphic and lithological data. These data have been used in conjunction with the gamma ray and well top data from Beswetherick (1995) to produce isopach maps of the Lower and Middle to Upper Devonian sedimentary rocks shown in figures 5.6 and 5.7, respectively. These figures illustrate the generally constant thickness of the Lower and Middle to Upper Devonian succession throughout much of the basin, but, where thickness changes do occur the gradients can be moderate to high. Figure 5.6 shows that there is a large area in the present day central area of the basin where Lower Devonian deposits are absent. Figure 5.7 illustrates an area in the west of the basin where the Middle and Upper Devonian succession is absent, as well as localised areas where the measured thickness deviates abruptly from the gentle regional trend (Fig. 5.7). One such area is in the NW of the basin where a thin succession of Middle to Upper Devonian sedimentary rocks occurs, this area of reduced thickness is oriented N/S (Fig. 5.7). Another area where variations in the thickness of the Middle to Upper Devonian succession can clearly be seen is in the central/ western area of the basin, this has an E/W orientation. The possible reasons for these and other thickness variations of the Lower and Middle to Upper Devonian succession are discussed in sections 5.6 and 5.7.

5.2.4. *Sequence A: Internal character and distribution.*

Sequence A (SA) is interpreted to occur throughout the area where gamma log subsurface data are available (Fig. 5.4). An isopach map of SA is given in figure 5.8. This sequence immediately overlies the Silurian shales of the Tanezzuft Formation which are of Llandoverian age according to Palaeoservices (1994b). These shales are generally organic rich and have moderate to high gamma ray counts that can easily be identified on wireline data (Fig. 5.4). The lowermost units in SA have a variable gamma ray response and in some instances the contact between shales of the Tanezzuft Formation and the fine-grained units in the lower part of SA can be difficult to pick as observed in wells F1-NC115 and C1-NC115 (Fig. 5.4). In this example the lower units in SA comprise sedimentary rocks with a moderate to high gamma ray response, similar to the response of the shales in the underlying Tanezzuft Formation. In contrast, the lowermost units of SA in well C1-NC174 have a low gamma ray signature, and the contact between SA and the Tanezzuft shales is picked out by a sharp decrease in gamma ray counts (Fig. 5.4). The contact between these two formations can also be picked by a change in the character of the sonic log wireline trace (Fig. 5.9).

The gamma log response curve of SA contains higher frequency cycles that show upward increases or decreases in gamma ray counts (Fig. 5.4). These gamma ray responses are generally recognised to result from changes in the amount of radioactive elements in the succession, usually associated with variations in grain-size and organic composition (section 2.3.1). Gamma ray wireline data from SA in the central and eastern part of the central area of the basin generally have a blocky expression with moderate to low total gamma ray counts (Fig. 5.4); interpreted as a sandstone rich succession containing low amounts of organic material and fine-grained sediment. The gamma ray wireline data from SA in the northern area of the basin generally have an irregular high amplitude signature with moderate to high total gamma ray counts (Fig. 5.4); interpreted as a sandstone poor succession rich in organic material.

The overall thickness of SA increases from south to north (figures 5.4 and 5.8), whereas in the west of the basin there is an area where SA is absent that is oriented E-W or SE-NW and has a narrow extension towards the north (Fig. 5.8). Immediately adjacent to this N-S extension the isopach contours are closely spaced indicating rapid changes in the thickness of SA. There also appears to be an increase in the thickness of SA in the SW and eastern areas of the basin and an area in the SE where SA is absent (Fig. 5.8). The possible origins of these thickness variations are discussed in sections 5.6.2 and 5.7.1.

5.2.5. *Sequence stratigraphic interpretation of Sequence A.*

The gamma ray wireline signature of SA usually contains cycles that show a repeated decrease in total counts upwards. These cycles are interpreted as parasequences, representing coarsening-upward cycles (section 2.4.4). The parasequences in SA stack to form distinct packages within which there is an overall coarsening or fining-upward signature. These packages can be defined using the systems tract framework outlined in section 2.4. The systems tracts interpreted in SA are displayed in figure 5.4, which also illustrates the lateral variability in the thickness and occurrence of these systems tracts. The absence of abundant sedimentological data relating to the sedimentary rocks from which these gamma ray wireline curves originate means that detailed analysis is difficult. However, according to RRI (1997) there is a decrease in palaeo-water depth upwards from the base to the top of SA in well H27-NC115.

Sequence A has been subdivided into two parasequence sets (Fig. 5.4), parasequence sets A and B (**PA** and **PB** respectively), the internal characteristics of which vary (Fig. 5.4). The gamma ray wireline trace of **PA** has a progradational signature (Fig. 5.4), interpreted as HST deposits, while the wireline trace of the overlying **PB** has a retrogradational then progradational signature (Fig. 5.4), interpreted as TST deposits overlain by a progradational HST succession. The TST in

PB is not observed throughout the basin (Fig. 5.4), probably as a result of localised variations in sediment supply and the rate of creation of accommodation space. An example of this may be in well C1-NC174 where a 15 metre thick sandstone dominated succession is observed beneath **PA** and **PB**. Sidewall core data from this sandstone indicate a broadly similar palaeoenvironment and age to the sidewall core samples analysed from sedimentary rocks in **PA** in the same well (Fig. 5.4). As this basal sandstone is not observed in any other well in the basin, its occurrence and distribution is interpreted to be controlled by localised changes in the rate of sediment supply and creation of accommodation space, rather than representing a separate stratigraphic cycle. These factors are also interpreted to affect the occurrence, thickness and gamma response of the parasequence sets in **SA**, with regional trends commonly modified by localised changes in these factors.

The stacking of the parasequence sets in **SA** may indicate the presence of a sequence boundary on the top of **PA**. However, there are no data relating to the nature of this surface in any of the wells and therefore, its character and importance are poorly constrained. The abrupt increase in gamma log counts at the top of **PA** is interpreted to record a flooding surface resulting from a rise in relative sea level. Such events can occur a number of times during a relative sea level cycle and are not always found associated with a sequence boundary (Posamentier et al. 1990). However, the transition from HST deposits in **SA** to TST deposits in the lower part of **SB** does suggest the presence of a sequence boundary between the systems tracts, and therefore, between **SA** and **SB**. This sequence boundary need not necessarily result from a major fall in relative sea level because of the low angle basin slope of the Murzuq Basin region during the Devonian. The relative sea level fluctuations associated with the transition from **PA** to **PB** are interpreted to be moderate to low, probably resulting from a high frequency, low magnitude fluctuation in relative sea level.

5.2.6. *Sequence B: Internal character and distribution.*

Sequence B (SB) is interpreted to be occur throughout the area where gamma ray wireline data are available (Fig. 5.4). An isopach map of SB is given as figure 5.10. The contact between the upper units of SA and the lower units of SB is generally picked by a sharp increase in gamma ray counts (Fig. 5.4). However, the contact is more subtle in several cases and is picked where there is a change from an aggradational to retrogradational wireline signature as shown in well C1-NC115 (Fig. 5.4). The passage from SA to SB is interpreted to represent a change from the dominantly organic poor sandstones in the upper part of SA to organic rich fine-grained rocks in the lower part of SB, as confirmed by the mud loggers reports.

The gamma log response curve of SB contains a number of higher frequency cycles that show upward increases or decreases in gamma ray counts. The gamma ray wireline data from SB in the central and northern part of the area generally have an irregular high amplitude signature with moderate to high total counts (Fig. 5.4). This gamma ray signature is interpreted to result from the interbedding of fine and coarse-grained rocks, while the dominantly high gamma counts indicate mainly fine-grained rocks. The gamma ray wireline data from SB in the eastern part of the area generally have a blocky expression with moderate to low gamma ray counts (Fig. 5.4). These data are interpreted to indicate a sandstone rich succession containing low amounts of fine-grained and organic material.

The overall thickness of SB increases from south to north as seen in figures 5.4 and 5.10. The latter figure is an isopach map constructed from bulk thickness data from SB in the central part of the Murzuk Basin. The thickness changes observed in SB (Fig. 5.10) closely correspond to those seen in SA (Fig. 5.8), there is an area in the west of the basin where SB is absent, and also areas in the SW and east of the basin where the thickness of SB increases (Fig. 5.10). The area where SB is absent is smaller than for SA and the northward extension of this area shown in figure 5.8 is absent. Explanations of the possible origin of these thickness variations are given in sections 5.6.2 and 5.7.1.

5.2.7. *Sequence stratigraphic interpretation of Sequence B.*

The gamma ray wireline signature of SB generally contains cycles that show a repeated decrease in total counts upwards, interpreted as parasequences that represent coarsening-upward cycles (section 2.4.4). The parasequences in SB stack to form distinct packages within which there is an overall coarsening or fining-upward signature. These packages can be defined using the systems tract framework outlined in section 2.4. The systems tracts interpreted in SB are displayed in figure 5.4 which also illustrates the lateral variability in the thickness and occurrence of these systems tracts. A lack of data from the sedimentary rocks in SB, as noted for SA, makes the detailed analysis of the corresponding wireline data difficult. However, core data from well H27-NC115 indicate a decrease in water depth upwards in sedimentary rocks coeval with SB (RRI 1997).

Sequence B contains two or three parasequence sets (Fig. 5.4), parasequence sets C, D and E (PC, PD and PE respectively), the thickness and occurrence of which vary in the basin. The gamma ray wireline trace of PC has a retrogradational signature with dominantly high gamma ray counts that increase upward and are interpreted as TST deposits (Fig. 5.4). Spectral gamma ray data indicate that PC is capped by a maximum flooding surface (Fig. 5.5). Parasequence set C is succeeded by PD and PE, both of which have wireline signatures that decrease upward, and are interpreted as HST deposits separated by a flooding surface. The thickness, occurrence and character of the wireline response of PC, PD and PE varies throughout the basin (Fig 5.4). The mechanisms driving the changes in these factors, such as sediment supply and the rate of creation of accommodation space, are interpreted to vary locally in the basin.

The stacked HST's in PD and PE contain lithological variations according to mud logging data, suggesting different conditions prevailed during the deposition of each parasequence set. The contact between PD and PE may be a sequence boundary and flooding surface across which there is a major change in depositional processes, or a flooding surface which juxtaposes different facies. The contact between PC and PD may have lesser importance in sequence stratigraphic terms, representing the equivalent of a type 2 rather than a type 1 sequence boundary or a parasequence set boundary surface. This surface need not necessarily result from a major change in relative sea level because of the low angle of floor dip during this time (section 2.4.7). However, it is recognised that the limited data relating to the surface between **PC** and **PD** mean that its character and importance are poorly constrained.

5.3. Seismic data.

The majority of seismic data interpreted during this study are from concession NC174, which is situated in the central part of the Murzuq Basin, comprising twenty-two lines from the 1992 NC174 survey (Fig. 5.2). Several isolated lines from earlier surveys were also available.

The seismic lines are of limited use interpreting the internal structure of the Devonian succession due to its reduced thickness throughout much of the basin (figures 5.8, 5.9). On these lines the base Devonian reflector can be identified but no top Devonian reflector can be discerned and there are no internal seismic facies. Unfortunately, where the thickness of the Devonian succession increases, towards the N and SE of the Murzuq Basin the quality of the seismic data is greatly reduced, being derived from 1977 and 1985 surveys. In view of these limitations the NC174 seismic data have been primarily used to interpret and map the nature of the lower contact of the Devonian succession and also to locate, map, and interpret structural trends in the central area of the basin.

5.3.1. *Interpreting the Seismic expression of the Lower Devonian boundary.*

According to wireline logs, mud loggers reports, core and sidewall core information there is usually an abrupt lithological change between the organic rich, fine-grained Tanezzuft shales and the overlying sandstone rich Devonian succession. This lithological change shows an offset on sonic logs (Fig. 5.9) and thus a significant acoustic impedance contrast occurs at this boundary. Such contacts are commonly resolvable on seismic data. Although facies changes in the Devonian succession can result in the juxtaposition of fine-grained units across the Silurian/ Devonian contact, this surface can be traced on seismic sections across much of the basin (figures 5.11 and 5.12).

The Devonian succession in the central area of the basin is dated Middle Devonian by Paleo-Services (1994a; 1994b) and RRI (1997), with no evidence of Lower Devonian sedimentary rocks in the area covered by the subsurface data base. The base Middle Devonian reflector is usually parallel to reflectors in the underlying Silurian units. However, between shot points 840 and 940 on line NC174-91-165 a low angle unconformity showing toplap can be observed beneath the Devonian units (Fig. 5.11).

5.3.2. *Subsurface structural interpretation.*

The seismic database outlined above was used to map the main structural elements in the central part of the basin, initially to ascertain their influence on depositional processes during the Devonian. The orientation and distribution of the main structural features identified there are displayed in figure 5.12. This figure also includes a top Silurian/base Mid Devonian TWTT isochron map. The isochron contours are affected by the faults, indicating post-Devonian movement on these structures. The variety of structures observed, generally strike N-S and are interpreted to form a regional scale, positive flower structure in concession NC-174 (S. Beswetherick 1995, pers comm.).

The structures in this central area of the basin comprise high angle faults with associated drape folds which in their present day orientation show reverse displacement (figures 5.13 and 5.14). The faults are interpreted to have a history of polyphase movement affecting the thickness of Palaeozoic strata (Fig. 5.14), with the latest deformation event resulting in strike slip displacement (T. Glover 1998, pers comm.). A detailed analysis of these structures is outside the scope of this study. However, an overview of the regional structural trends is given in this chapter and details are provided in Glover (1999).

5.4. Integration of subsurface and outcrop data.

The outcrop data from the Lower to Upper Devonian sedimentary rocks presented in chapters 3 and 4, and the subsurface data outlined above, have been integrated to create a regional sedimentological and subsidence model for the Murzuq Basin. This model includes published data from the Djado and Dur al Qussah sub-basins to the south and east respectively, as well as from the western and eastern margins of the basin (Fig. 1.3). It should be remembered that during the Palaeozoic, specifically the Devonian, the Murzuq Basin did not exist as a separate depositional basin, but was situated on the northern margin of Gondwana. During the following sections all references to specific areas in the Murzuq Basin relate to their location in the area delimited by the present day outline of the basin.

5.4.1. *Early Devonian palaeogeography and regional distribution.*

As described in chapter 3 and sections 5.1 to 5.3.1, sedimentary rocks of Early Devonian age are well exposed on the western margin of the basin, while they are absent from much of the central area of the basin (Fig. 5.6). On the western margin of the basin the Lower Devonian Tadrart Formation comprises fluvial to shallow marine facies (figures 3.15, 3.20 and 3.23), which are overlain by the shallow marine sedimentary rocks of the Ouan Kasa Formation (Fig. 3.28). Previous work by Clarke-Lowes (1985) and Massa (1988) in the southern Murzuq basin and Beuf et al. (1971) and Mergl and Massa (1998) in the Djado sub-basin (Fig. 1.3), showed that the Lower Devonian succession comprises continental facies that are overlain by shallow marine sedimentary rocks. The Lower Devonian succession also crops out along the eastern margin of the basin. Bellini and Massa (1980) interpreted the Lower Devonian Tadrart Formation in this region as braidplain deposits, overlain by marginal marine deposits that are succeeded by the marginal marine Ouan Kasa Formation. The angular contact between the Silurian Akakus Formation and the Lower Devonian Tadrart Formation west of Al-Awaynat (Fig. 3.6) observed during this study, and by previous workers, notably Bellini and Massa (1980), indicates a geographically localised phase of pre-Early Devonian uplift (section 1.2.3).

According to Bellini and Massa (1980) on the eastern margin of the basin the thickness of the Lower Devonian succession is affected by localised palaeotopographic variations. These thickness changes are most notable in the Ouan Kasa Formation which is 100-200 metres thick on the eastern margin, compared to 40 metres thick on the western margin. In the SE of the region, near the Mourizidie horst (Fig. 1.3), the entire Devonian succession is condensed, relative to the thick succession in the Dur-al Qussah sub-basin to the north (Bellini and Massa 1980).

The subsurface data outlined in sections 5.1 to 5.3.1 show that the Lower Devonian succession is absent from much of the central area of the basin (Fig. 5.6), which can be explained by non-deposition, erosion or a combination of the two. The Silurian Akakus Formation is also absent from this central region of the basin, in contrast to the SW margin of the Murzuq Basin where it is 250 metres thick (Fig. 1.3). On the SW flank of the Murzuq Basin the Tadrart Formation thins northward (Fig. 3.29). Close examination of the Tadrart Formation shows that both the fluvially dominated lower Tadrart and the marginal marine upper Tadrart facies thin northward. These data suggest a long-lived decrease in the rate of creation of accommodation space towards the north (Fig. 5.15), rather than the onlap of a pre-Early Devonian palaeotopography. The northward decrease in subsidence interpreted in Lower Devonian sedimentary rocks on the western margin of the basin also appears to have had little or no effect on fluvial drainage patterns which were consistently towards the NW. Palaeocurrent data from the Tadrart Formation indicate fluvial drainage from SE to NW with the palaeoshoreline oriented NE-SW. These data suggest that the present day central area of the basin was not a major topographic high that influenced the drainage pattern of coeval fluvial systems. If substantial uplift of the central region occurred then the regional northwesterly slope of the basin area would be modified, affecting the drainage directions of these fluvial systems. However, there is no evidence that the direction of fluvial drainage was significantly modified during this time. The rapid pinch-out of the Tadrart Formation also occurs in an easterly direction towards the present day central area of the basin (Fig. 5.16), interpreted to be the result of a NW striking fault zone situated to the west of NC-174

(Fig. 5.17). The exact position of this structural zone is difficult to ascertain. However, the presence of a major structure in this area of the basin has been proposed a number of oil company workers, including M. Eales (1988, pers comm.).

In the central area of the Murzuq Basin the Lower Devonian sedimentary rocks may be attenuated or absent due to non-deposition, rather than substantial post-depositional erosion (Fig. 5.18). Assuming region wide deposition during the Early Devonian, a phase of post-depositional erosion would need to have removed a full succession of Lower Devonian sedimentary rocks from the central area of the basin. The entire Akakus Formation and upper part of the Tanezzuft Formation are also missing from this region, which would therefore indicate the erosion of over 500 metres of a sandstone dominated Silurian/Devonian succession. An erosive event such as this would probably have been driven by fluvial incision and aggressive subaerial weathering in moderate to high energy environments. As the sub-Devonian Tanezzuft Formation is predominantly shaley, any such high energy erosive events during the late Silurian to Early Devonian would almost certainly cause deep angular incision of these fine-grained units. However, such erosive features are not observed on seismic sections in this central region of the basin. For these reasons the absence of the Lower Devonian and part of the Silurian succession is not thought to be the result of the post-depositional erosion of an originally complete succession.

The central area of the basin may have had a thin Lower Devonian succession deposited in it, most likely deposited during the phase of relative sea level rise interpreted in this study to have occurred at the end of the Early Devonian (section 3.9). A minor erosive episode in this central part of the basin would have been sufficient to remove this thin Lower Devonian succession. Such an erosive event is locally recognised during the late part of the Early Devonian (Emsian) on the western flank of the basin, removing the Ouan Kasa Formation (D. Massa 1998, pers comm.).

This study interprets the absence of the Lower Devonian sedimentary rocks from the present day central area of the basin to be a result of a combination of low rates of creation of accommodation space (relative to the SW margin), and a minor phase of post depositional erosion. This central area of the basin contains a number of seismically resolvable faults outlined above. Strike-slip movement of these faults may have resulted in differential subsidence during the Early Devonian and caused thickness changes in the underlying Palaeozoic succession (T. Glover 1998, pers comm.). These variations in subsidence do not appear to be associated with major re-orientations of the palaeoshoreline and up-dip fluvial systems during the Early Devonian. Sedimentological data from Lower Devonian succession throughout the basin suggest that the palaeoenvironmental conditions for the Early Devonian outlined in chapter 3 prevailed across the entire Murzuq Basin region, i.e. continental at the base of the Early Devonian, marginal marine in the upper part of the Early Devonian.

On the SW margin of the Murzuq Basin, during deposition of the lower to middle parts of the Tadrart Formation shorter term cycles in relative sea level are identified. These relative sea level fluctuations resulted in the formation of seven depositional sequences (S1 to S7) that contain and/or are separated by sequence boundaries and flooding surfaces (section 3.9). During this time the palaeoshoreline only rarely impinged upon the Murzuq Basin region. Given the low angle slope of the basin this suggests the outcrop study area on the SW flank of the basin was situated many tens or hundreds of kilometers up-dip of the palaeoshoreline.

During deposition of the upper part of the Tadrart Formation shallow to marginal marine conditions prevailed across most of the basin as a result of relative sea level rise (Fig. 3.25). As stated in sections 3.5 and 3.6, during this time large fluvially dominated braid deltas occurred throughout the region. These delta deposits were reworked by minor amounts of wave action, creating small shoreface complexes. Although the position of the palaeoshoreline during this time is difficult to constrain, sedimentological data suggest a position near or slightly up-dip of the outcrop studied on the SW flank of the basin. The rise in relative sea level responsible for the deposition of the upper part of the Tadrart Formation is interpreted to have continued, resulting in a major shut-off of siliciclastic sediment supply and deposition of the shallow marine, oolitic Ouan Kasa Formation across much of the basin. This change in depositional style was also broadly coeval with a phase of late Early Devonian tectonism which may also have affected the sediment supply processes.

5.4.2. *Mid to Late Devonian palaeogeography and regional distribution.*

As described in chapter 4 and sections 5.1 to 5.3, Middle and Upper Devonian facies are well exposed on the northern margin of the basin, and occur throughout much of the basin according to subsurface data (Fig. 5.7). This stratigraphic interval has been subdivided into six sequences (S8 to **S13**), within which eight parasequence sets (**PS1** to **PS8**), parasequences and key sequence stratigraphic surfaces have been identified. A comparison of the outcrop and subsurface subdivisions of the Middle and Upper Devonian succession can be seen in figure 5.19. Previous work by Protic (1984) and Massa (1988; 1992; and 1998) in the south and west of the basin identified marginal to open marine facies of Mid to Late Devonian age in this region. The more northern of these two areas is in the Tikiumit map sheet area of Protic (1984), also known as the Talagrouna type section of Massa (1992) (Fig. 5.20). In this locality the sedimentary rocks are dated Late Devonian (upper Frasnian) by the fauna (Mergl and Massa 1992), and are interpreted as coeval with the upper part of **S10** or **S11** on the southern flank of the Gargaf Arch.

According to Protic (1984) lenses of the Middle to Upper Devonian succession dip at between 20°and 36° towards the NE near the Tikiumit dislocation. The Tikiumit dislocation is interpreted to be part of the Tihemboka Arch, or Tihemboka uplift/anticline of Klitzsch (1981) and Goudarzi (1980) respectively. According to Goudarzi (1980) the structure was uplifted during the Mid Devonian; this phase of movement is also recognised in this study. However, the timing of the subsequent tilting of the Mid/Late Devonian and early Carboniferous units in the Tikiumit area recognised by Protic (1984) is difficult to constrain due to limited data. Movement of the Tikiumit dislocation and Tihemboka Arch may have occurred as recently as the Mesozoic (T. Glover 1998, pers comm.). The absence of Middle/Upper Devonian deposits that make up all, or part of, **S8**, **S9**, and also the lower part of **S10** and **S11** in the Tikiumit region, may be due to non-deposition or post-depositional erosion. However, the presence of the deposits comprising the upper part of **S10**, **S11** or **S12**, which are recognised to be strongly transgressive elsewhere in the basin, may indicate the onlap of palaeo-highs during a rise in relative sea level. The onlap of one such region is shown in figure 5.21, where conditions of fluctuating relative sea level result in separate periods of deposition and erosion or exposure on a topographic high.

Middle to Upper Devonian sedimentary rocks also occur south of the Tikiumit area and were studied in the Gour Iduka region by Massa (1988) (Fig. 1.3) and Mergl and Massa (1992) (Fig. 5.19). According to these authors the succession in the Gour Iduka region comprises sedimentary rocks of the Awaynat Wanin Group that can be correlated with coeval facies elsewhere in the basin (Fig. 5.22). The stratigraphic thickness of the Awaynat Wanin formations I and II (S8 and S9) in the Gour Iduka region is broadly comparable to that observed on the southern flank of the Gargaf Arch, indicating similar subsidence/ accommodation space histories (Fig. 5.19). The reduced thickness of the Awaynat Wanin Formation III (S10), and reduced thickness or absence of Awaynat Wanin formations IV and V (S11 to S13), could be due to non-deposition or post-depositional erosion. However, the absence or attenuation of these Upper Devonian facies in the Gour Iduka region was probably due to this area being up-dip of any relative sea level fluctuations, and primarily an area of sedimentary by-pass and erosion.

Sedimentary rocks of Mid to Late Devonian age also crop out in restricted areas of the eastern flank of the basin where they comprise a 150-300 metre thick succession of marginal marine sandstones (Bellini and Massa 1980; Mergl and Massa 1992). The Middle to Upper Devonian sedimentary rocks in this region can be correlated with coeval facies in the upper part of S10, S11 or the lower part of S12 on the southern flank of the Gargaf Arch (Fig. 5.22). Bellini and Massa (1980) also interpreted this eastern flank of the basin as an area of instability, resulting in the absence of part of the Middle to Upper Devonian succession equivalent to S8, S9 and the upper part of S12. These authors also interpreted that the increased thickness of the Devonian succession in this region to be the result of this instability.

The six sequences interpreted in the Middle to Upper Devonian succession on the southern flank of the Gargaf Arch, and described in chapter 4 (Fig. 4.91), are interpreted to have formed as a result of eustatically driven changes in relative sea level (section 4.7.3). In the central area of the basin wireline log data have been used to identify two sequences in the Middle to Upper Devonian succession (Fig. 5.3). These two sequences also contain a number of parasequence sets formed as a result of higher frequency and lower magnitude fluctuations in relative sea level (figures 5.3 and 5.4). The surfaces bounding these parasequence sets can be correlated with similar ones identified in outcrop (section 4.7) and are detailed below and in figure 5.22.

Palynological data from SA indicate an Eifelian/Givetian age for the lower part and a Givetian age for the upper part of the sequence in the central part of the basin (Palaeoservices 1994a; 1994b). These age constraints suggest that SA may be coeval with some or all of PS1, PS2, PS3 and PS4 (S8 and S9; the B'ir al Qasr and Idri formations respectively) (Fig. 5.22). Palaeoenvironmental interpretations of the B'ir al Qasr and Idri formations identify an increase in palaeowater depth, and therefore a rise in relative sea level, at the top of PS3, that is of greater magnitude than those associated with the upper boundary of PS1 and PS2. This increase in relative sea level resulted in a change from sand to shale dominated deposition, which would normally give a sharp decrease in gamma ray counts. Such a surface is located in SA at the contact between PA and PB and is thought to correlate with the upper surface of PS3. Therefore, the lower part of SA is equivalent to PS1, PS2 or PS3, while the upper part of SA is equivalent to PS4 (Fig. 5.22). It may be possible to more accurately correlate between these sequences as there is more evidence of marine influence in PS3 when compared to PS1 and PS2. It is therefore possible that PS3 correlates with the lower part of SA, although the paucity of detailed subsurface data makes definitive correlation difficult.

The lack of stratigraphic age data from SB makes direct correlation with outcrop difficult. However, the three parasequence sets identified in SB are thought to correlate with the parasequence sets observed in outcrop on the southern flank of the Gargaf Arch. The lower parasequence set (PC) is thought to correlate with PS5 and PS6 (Fig. 5.22), although the pinch out of PS5 on the southern flank of the Gargaf Arch may be repeated in the subsurface. The boundary between PS5 and PS6 is difficult to recognise in PC using wireline data. However, it is interpreted to occur where the pattern of the wireline trace becomes retrogradational (Fig. 5.4), the transition representing a flooding event that caps PS5. The flooding surface is important as it marks a surface across which there is a major change in depositional processes from a dominantly siliciclastic succession to one dominated by oolites. This change in depositional processes is interpreted to result from a rise in relative sea level responsible for a major increase in palaeo-water depth (section 4.7.3). Parasequence set C is therefore capped by a flooding surface that correlates either with the boundary between S11 and S12, or occurs within the lower part of S12. This flooding surface is characterised by high total gamma ray counts, and in well A1-NC174, high amounts of uranium (section 5.4.2, Fig. 5.5). Outcrop studies of the eastern margin of the basin by Bellini and Massa (1980), and the western margin of the basin by Bellini and Massa (1980); Massa (1988) and Mergl and Massa (1992; 1998), noted a succession of sedimentary rocks coeval to PS5 and/or PS6, directly overlying pre-Upper Devonian deposits. This distribution of PS5 and PS6 is interpreted to indicate the onlap of palaeohighs as a result of relative sea level rise.

Parasequence sets D and E have shallowing-upward signatures according to wireline data (section 5.4) and are correlated with PS7 and part of PS8 that also have shallowing-upward trends. The upper limit of PE is picked as the Devonian/ Carboniferous boundary which represents the base of a regressive (transgressive) succession of Carboniferous (Tournaisian) sedimentary rocks (Fig. 5.4). The nature and exact position of this boundary in the subsurface is poorly defined due to limited data. However, the upper surface of PE is probably diachronous in view of its origin as a flooding surface, as the timing of transgressive events are strongly affected by localised changes in sediment supply and the rate of creation of accommodation space. Outcrop studies by Parizek et al. (1984) and Seidl and Rohlich (1984) identified the Ashkidah Formation (PS8) as a transitional Devonian/ Carboniferous age succession. Therefore the separation of PE from the overlying Carboniferous facies is somewhat arbitrary in chronostratigraphic terms.

Palaeogeographic reconstructions based on outcrop data from the southern flank of the Gargaf Arch for the Mid to Late Devonian can be seen in figures 4.17, 4.26, 4.42, 4.47, 4.57 and 4.68. Analysis of core data by RRI (1997) from the central area of the Murzuq Basin (NC115) recognised a number of fine-grained deltas with the coeval palaeoshoreline oriented E/W (Fig. 5.23). These deltas in the central area of the basin prograded from south to north (RRI 1997). In areas proximal to the Gargaf Arch region the orientation of the palaeoshoreline, and associated deltas, is interpreted to swing around towards the NE-SW and N-S as shown in figures 4.17, 4.26, 4.42, 4.47, 4.57 and 4.68. Sidewall core data from the central area of the basin show abundant marine microplankton and miospores (Paleo-Services 1994a; 1994b) indicating the presence of shallow marine environments receiving abundant amounts of terrestrially derived kerogen. These variations in palaeoshoreline orientation are interpreted to result from differential subsidence throughout the basin.

The relative sea level cycles interpreted in the Middle to Upper Devonian succession on the southern flank of the Gargaf Arch have a variable magnitude and amplitude (Fig. 4.76). As these relative sea level fluctuations occurred the palaeoshorelines at the time will have moved up or down the dip slope of the basin. The distances involved in these palaeoshoreline movements are difficult to constrain but may have been substantial (section 2.4.6). The regional implications for these relative sea level fluctuations are discussed in section 6.5.2.

The onlap of palaeohighs in the Murzuq Basin during the Late Devonian (Frasnian) transgression (PS5 and PS6) may indicate a moderate amplitude rise in relative sea level. However, the absence of Middle to Upper Devonian deposits on the Awaynat uplift may indicate that the majority of the transgressive events affected widely spaced and up-dip areas as a result of the low angle basin slope, rather than the high amplitude of these transgressions. The absence of sedimentary rocks deposited in S8 to the lower part of S10 from the eastern and western flanks of the basin indicates either primary non-deposition, or in view of their proximity to large structures, post-depositional erosion (Fig. 5.21). The onlap of facies deposited in the upper part of S10 to the lower part of S12 onto these two palaeohighs indicates that the relative sea level rise associated with their deposition was of sufficient magnitude to onlap part or all of their subaerial relief. Examination of facies in the present day central area of the basin comprising the lower part of **PD**, and coeval outcrop sedimentary rocks in the lower part of **PS7**, reveals that this relative sea level rise was not only of moderate magnitude, but also of regional extent.

5.5. A summary of the structural activity in the Murzuq Basin during the Devonian.

The regional thickness trends of the Lower to Upper Devonian sedimentary rocks in the Murzuq Basin described in sections 5.1 to 5.4, and outlined in figures 5.6 and 5.7, are interpreted to be influenced by tectonism. This tectonic activity caused the movement of a number of structures that can be identified in the basin using outcrop and subsurface data (figures 5.24, 5.25 and 5.26). Faults analogous to the N and NE striking, large scale and closely spaced faults observed in the central area of the basin were not observed in outcrop on the west or northern flanks of the basin, and are not reported on the eastern flank. The vast majority of the structures observed on the southern flank of the Gargaf Arch are E-W silicified fault zones (T. Glover 1998, pers comm.). Geological maps of the SW margin of the Murzuq basin depict a N-S and NW-SE structural trend in the basement as in the central part of the basin. However, the drape folds associated with these faults in the subsurface are not observed in outcrop. Conversely, the 10°30' fault and Tihemboka Arch are much more extensive than any of the structures resolved on seismic from the central area of the basin. These data suggest that during the Early to Late Devonian the effects of tectonism varied from one part of the basin to another, probably as a result of the varying rheological properties of different parts of the basement. These structures were active at various times during the Devonian and are detailed in the following sections.

5.5.1. *A summary of the structural activity during the Early Devonian.*

The thickness and distribution of the Lower Devonian succession in the Murzuq Basin are affected by pre-Devonian and end Early Devonian tectonism. The structures observed on seismic sections (Fig. 5.12), and those interpreted on the margins of the basin in this study and by previous workers (Fig. 5.25), are considered to have been active throughout the Palaeozoic, including the Early Devonian. The majority of these structures are interpreted as basement penetrating, steep, reverse faults oriented N/S or NW/SE (figures 5.11, 5.12, 5.13, 5.14 and 5.25).

The absence of Lower Devonian deposits from the central and west central areas of the basin (Fig. 5.6) is interpreted to be the result of a decrease in the rate of creation of accommodation space in these regions (Fig. 5.15). As shown in figures 5.12 and 5.25, these areas contain a number of major structures that may have been active during the Early Devonian. The western margin of the region, where the Lower Devonian succession is absent, is thought to correspond with the Awaynat uplift (Fig. 5.25), a structure that is interpreted to have influenced sedimentation during most of the Early Devonian (Fig. 5.27). An example of the pre-Lower Devonian fault activity can be seen on the western margin of the basin near Al-Awaynat where units of the Akakus Formation are tilted beneath the Lower Devonian Tadrart Formation (Fig. 3.6). The 10°30 fault zone may be responsible for this phase of deformation (Fig. 5.25), itself part of the larger Tihemboka Arch structure shown in figures 5.25 and 5.26.

Previous work in the Murzuq Basin has also recognised a phase of uplift and erosion during the late Silurian (Bellini and Massa 1980; Clarke-Lowes 1985), resulting in a depositional hiatus between the Akakus and Tadrart formations. The uplift of the Akakus Formation on the western margin, and removal of the Akakus and part of the Tanezzuft formations from the central area of the basin are interpreted to be the result of this phase of tectonism (Fig. 5.27). The Tanezzuft Formation is also absent from the Brak Bin Ghanimah Arch, most notably on the southern flank of the Gargaf Arch where Middle Devonian and Cambro-Ordovician sedimentary rocks are juxtaposed (Fig. 4.2a). Another phase of deformation began at the end of the Early

Devonian (Emsian), resulting in the uplift of the western margin of the basin near the 10°30' fault zone, Awaynat uplift and Tihemboka Arch, and on the eastern margin near the Mourizidie Horst (Fig. 5.27).

The origins of these phases of tectonic activity during the Early Devonian are poorly understood. It is reasonable to assume that several phases of deformation occurred, possibly with extensional and contractional deformation during this time, with intermittent strike-slip movement (T. Glover 1998, pers comm.; Glover 1999). As these phases of tectonism are poorly constrained, the orientation of the maximum principal stress ($\sigma 1$) during these events is difficult to estimate. The heterogeneities recognised in the underlying basement across North Africa may also have influenced the nature and timing of fault activity during deformation (Glover 1999).

Palaeocurrent data from the western margin of the Murzuq Basin prove that the Tihemboka Arch was not active during much of the Early Devonian and that the basin floor dipped at a low angle towards the NW (Fig. 1.12). Analysis of the Tihemboka Arch reveals a complex depositional, subsidence and uplift history during the Devonian, with the northern part of the structure possibly influenced by the Awaynat Uplift (figures 5.25 and 5.26).

5.5.2. *A summary of the structural activity during the Mid to Late Devonian.*

The regional thickness trends of Middle to Upper Devonian units in the Murzuq Basin outlined in sections 5.1 to 5.3.1 are influenced by tectonic activity. The end Early Devonian tectonic event outlined above terminated during the early part of the Eifelian when deposition of SA resumed across much of the basin (Fig. 5.7). The lower part of the Mid Devonian appears to be a period of tectonic quiescence and there is a clear shift in the spatial distribution of areas uplifted during the pre-Early Devonian tectonic event compared to those uplifted during the Emsian/Eifelian event (figures 5.25 and 5.27). The southwestern area of the basin becomes an area of uplift or non-deposition while conditions of relative sea level rise are noted in the central area of the basin, an Early Devonian palaeohigh. The structures observed on seismic sections in the central area of the basin are not seen to influence the thickness or facies of the Middle to Upper Devonian succession, possibly because the succession is too thin to record such delicate changes. An exception to this may be found near the Serdeles fault zone (Fig. 5.25) where a reduced thickness of SA and SB is probably the result of a fault controlled decrease in the rate of subsidence. On the southern flank of the Gargaf Arch the thickness of Middle to Upper Devonian sedimentary rocks decreases towards the east (Fig. 4.78), most notably in the Quttah Formation (PS5) which is entirely absent from the east/central region of the Gargaf Arch outcrop. These thickness changes are interpreted to be the result of movement of the Brak Bin Ghanimah Arch during the Frasnian, comprising the southern part of the Tripoli-Tibesti Uplift (Fig. 1.6), which led to the uplift, attenuation and erosion of PS5 on the crest of the structure. This Frasnian tectonic activity may be one of the mechanisms that led to a change from coarse-grained siliciclastic to fine-grained oolitic deposition as shown in figure 4.74, modifying the basin profile, sediment supply processes and the regional palaeoenvironment. The onlap of Upper Devonian sedimentary rocks onto the flanks of structures on the eastern and western margins of the basin are interpreted to result from relative sea level rise during a period of tectonic quiescence and regional subsidence (figures 5.21 and 5.27). The absence of Middle Devonian sedimentary rocks from the margins of these structures may indicate that they influenced subsidence rates even if the structures themselves were inactive.

5.6. Palaeoclimate and sediment source areas during the Devonian.

As considered in chapters 1, 3 and 4, the palaeoclimate during the Early to Late Devonian is difficult to constrain. The occurrence of Lycophytes and miospores in the Lower to Upper Devonian succession illustrates the colonisation of some marginal marine to continental areas by plants. However, little work has been done to evaluate their use as palaeoclimatic indicators. The absence of palaeosols documented or observed in this study in the Devonian succession also removes another possible source of palaeoclimatic data.

Palaeocontinental reconstructions position the Murzuq Basin in low to mid latitudes during the Early Devonian (Fig. 1.13), within the present day temperate zone. At that time the land surface was similar to that observed in present day arid climates due to the limited vegetation cover (Schumm 1969; RRI 1989; Miall 1996) (Fig. 1.13). Those conditions prevailed across the entire basin up-dip of the bayline, with a large braidplain present in up-dip areas (figures 3.15 and 3.23). There is no evidence that any areas in the Murzuq Basin had substantial topographic relief, although small areas in the Tibesti Massif region may have acted as localised sources of sediment, at or near the Mourizidie horst (Fig. 5.25). The movement of the Tihemboka Arch at the end of the Early Devonian (Emsian) does not appear to have affected palaeocurrents during this time, and the arch seems not to have acted as a major source of sediment during deposition of the subsequent Mid to Late Devonian succession. However, at the end of the Early Devonian the Tihemboka Arch was uplifted sufficiently to allow erosion of the Lower Devonian Tadrart and Ouan Kasa formations, and also part of the Silurian succession, from the crest of the structure (Fig. 5.26).

During the Mid to Late Devonian Gondwana drifted southwards and the Murzuq Basin moved further into mid-latitudes (Fig. 1.11). There is a decrease in the sand-shale ratio in the Middle to Upper Devonian succession on the southern flank of the Gargaf Arch and in the subsurface. This change in grain-size may indicate a decrease in depositional energy, or reflect changes in weathering processes, sediment source character and depositional energy associated with palaeoclimatic conditions. Variations in the character of sediment supply processes may indicate a tectonically driven change in basin physiography at the end of the Early Devonian, an expression of which may the uplift of the Tihemboka Arch (Fig. 5.26). This may then have been emphasised by the tectonic activity identified at the end of the Frasnian (end PS5) which further reduced the amount of siliciclastic sediment supplied to much of the basin.

Chapter 6. A comparison of the structural and sedimentological evolution of the Murzuq area with other coeval sedimentary accumulations.

6.1. Introduction to the Devonian stratigraphic evolution.

The structural setting, sedimentology and sequence stratigraphic framework for the Devonian succession in the Murzuq Basin detailed in chapters 3, 4 and 5, are compared and correlated with those from coeval facies in North Africa and other areas around the world. Information from outside the Murzuq Basin area comprises well logs, oil company reports and field guides as well as previously published work. This information is synthesised in sections 6.2 to 6.4 and correlated with the interpretations made from coeval sedimentary rocks in the Murzuq Basin in sections 6.5 and 6.6.

6.2. The distribution and character of Devonian facies in the Afro-Arabian areas.

During the Devonian the region now occupied by the Murzuq Basin was situated on the northern margin of Gondwana (figures 1.10 and 1.11). This region was subdivided into separate cratonic 'sag' basins during post-Devonian time (Fig. 1.2). Data from the Devonian sedimentary rocks that crop out on the margins of a number of these basins, as well as oil exploration data relating to coeval units in the subsurface areas of these basins, are discussed below.

6.2.1. *Ghadames Basin: Early Devonian stratigraphic evolution.*

The Ghadames Basin in northern Libya, eastern Algeria and Tunisia (Fig. 1.2) contains Lower Devonian sedimentary rocks according to subsurface data. A stratigraphic column of these rocks in the Ghadames Basin can be seen in figure 6.1.

<u>Libyan Ghadames Basin</u>

In the southern part of the Ghadames Basin the Lower Devonian Tadrart Formation overlies progressively older sedimentary rocks southward (unpublished ECL report 1989). The basal Tadrart Formation comprises polymict conglomerates or gravels (ECL 1989; Bekkouche 1992), passing up into poorly to moderately sorted planar and trough cross-bedded sandstones (ECL 1989). The sedimentary rocks of the Tadrart Formation are overlain by the green/brown claystones and thinly bedded sandstones of the Ouan Kasa Formation (ECL 1989; Gundobin 1985). In the northern part of the Libyan Ghadames Basin the Lower Devonian succession comprises Pragian coarse-grained sandstones and carbonates and Emsian sandy carbonates (Bellini and Massa 1980). The Pragian sandstones overlie rocks of uppermost Silurian (Pridoli) age (Bellini and Massa 1980). These Early Devonian sedimentary rocks are coeval with the Tadrart and Ouan Kasa formations respectively in the Murzuq Basin that are described in chapter 3 (sequences 1 to 7). An example of the gamma ray and Neutron log response of the Lower Devonian succession in well D1-66 can be seen in figure 6.2, illustrating the blocky wireline signature of the Tadrart Formation compared to the irregular high amplitude signature from the Ouan Kasa Formation.

In the central and northern part of the Ghadames Basin the Tadrart Formation comprises rhythmic fining-upward cycles, interpreted as shallow marine facies influenced by strong tidal currents (ECL 1989), while Bracaccia et al. (1991) inferred periods when wave action dominated. In the northern part of the Libyan Ghadames Basin the Lower Devonian Tadrart and Ouan Kasa formations form a thick transgressive sequence, comprising fluvial sandstones in the lower Tadrart Formation, passing upward into shallow marine shales, sandstones and carbonates in the upper part of the Tadrart and Ouan Kasa formations (Poyntzt 1995). A shallow marine palaeoenvironment is also interpreted for the sedimentary rocks in the upper part of the Tadrart Formation in the northern part of the Ghadames Basin on the basis of core data (Bracaccia 1991). In the Ghadames Basin the Lower Devonian Tadrart Formation was divided into nine members, deposited during the Gedinian to early Siegenian by Bekkouche (1992). This author also recognises similar cycles in the Ouan Kasa Formation, interpreting seven sub-units within the Ouan Kasa Formation in the Libyan Ghadames Basin.

Work by Belhaj (1998) interpreted the Lower Devonian succession in the Ghadames Basin as continental, shallow marine and inner neritic deposits, with the

Tadrart clastics sourced from the Gargaf Arch and from poorly defined 'more northerly and southern regions'. However, the Gargaf Arch is not recognised in this study to be a major source of sediment during the Early Devonian, as palaeocurrent data from Cambrian to Late Devonian sedimentary rocks of the Murzuq Basin, and work by Glover (1999), indicate that this structure did not form until after the Devonian (chapter 5). During the Early Devonian the central part of the Murzuq Basin was above regional base level (Fig. 5.6), but most clastic sediment was derived from areas to the S or SE of this region (section 1.4.1). A southerly source of sediment in the Ghadames Basin in the Early Devonian is recognised by Belhaj (1998), but he suggested that the lower part of the Tadrart Formation was derived from a northerly source.

The Ouan Kasa Formation is interpreted by Braccacia et al. (1991) to be a separate sedimentary cycle in the northern part of the Ghadames Basin with an emergence surface marking its lower contact with the Tadrart Formation. Iron-oxide rich sandstones occur in the Ouan Kasa Formation and at its base, the latter units marking the transition between the Tadrart and Ouan Kasa formations. These iron-oxide rich sandstones were interpreted by Bracaccia et al. (1991) as sub-aerial exposure surfaces formed at the end of sedimentary cycles.

According to Massa (1988), the Lower Devonian succession in the Ghadames Basin thickens northward although localised lateral variations in thickness are also present (Fig. 6.3). The possible causes of these thickness variations are discussed in sections 6.5.1 and 6.6.1.

Algerian Ghadames Basin

A phase of late Silurian to Early Devonian tectonism is interpreted to have uplifted several areas, including the Hoggar-Reguibat massifs, the Tihemboka Arch, the Ahara Arch, and the High Plateau in the north of Algeria and Tunisia (RRI 1989; Anadarko 1993a) (Fig. 1.2). These areas acted as localised clastic sources which were transgressed during the Gedinnian/Siegenian (RRI 1989; Anadarko 1993a). In the Algerian part of the Ghadames Basin (Fig. 1.2) the Lower Devonian sedimentary rocks comprise sandstones and siltstones with a high degree of lateral facies variability (Alem and Benebouche 1995). The separation of Lower Devonian and Silurian sedimentary rocks in the subsurface can be problematic due to their similar response on wireline logs and the concordant nature of their contact (Alem and Benebouche, 1995). An example of the wireline log response of the Lower Devonian succession in the Algerian Ghadames Basin can be seen in figure 6.4.

The Lower Devonian succession in Algeria is usually grouped with the Upper Silurian sandstones into the F6 reservoir sequence, containing the M1, M2, A, B1, B2, C1, C2 and C3 sub-units (Fig. 6.1). These sedimentary rocks are interpreted as shelf sandstones and shales (Anadarko 1996a; 1996b) or fluvial, aeolian, coastal plain and shoreface deposits (Bekkouche 1992) in the far north of the basin; tidal channel and offshore bar sedimentary rocks in the central part of the basin (Alem and Benebouche 1995) and meandering fluvial deposits in the south of the basin (Bekkouche 1992; Anadarko 1993). The Lower Devonian sedimentary rocks in the Ghadames Basin are inferred to be lowstand incised valley fill deposits in the south of the basin and lowstand wedge deposits in the north of the basin, becoming transgressive in the upper part of the succession (Bekkouche 1992). The transgressive nature of the upper part of the Seigenian succession was also recognised by Scott (1994), who interpreted a thick succession of Siegenian sedimentary rocks to have been deposited during a number of transgressive Systems Tracts, forming a backstepping sequence set. Bekkouche (1992) recognises the boundary between the Siegenian C3 and Emsian F4 deposits as transitional, with the F4 and F5 sandstones and shales deposited within Transgressive and Highstand Systems Tracts (Fig. 6.1) (section 6.6).

The Lower Devonian succession is absent from the crest of the Ahara Arch on the southern flank of the basin, and is attenuated in wells on the northern flank of this structure (Ford and Muller 1995; Anadarko 1993b ; 1996a). These thickness changes are interpreted to be the result of erosion at the end of the Early Devonian or during the Frasnian (Late Devonian) (Anadarko 1996a; 1996b; Ford and Muller 1995; L. Villa Lobos 1996, pers comm.). A summary of the unconformities interpreted by

Anadarko (1993a; 1994; 1996a; 1996b) within the Devonian succession in the Algerian Ghadames Basin is given in figure 6.5.

6.2.2. *Ghadames Basin: Mid to Late Devonian stratigraphic evolution.*

Data relating to the Middle to Upper Devonian sedimentary rocks within the Ghadames Basin includes outcrop and subsurface data, the latter including information from well formation tops, well logs, core reports and previously published work.

Libyan Ghadames Basin

Well top data from the Libyan Ghadames Basin were used by El-Rweimi (1991) to produce an isopach map of the Middle to Upper Devonian Awaynat Wanin Group (Fig. 6.6). This isopach map shows an overall thickening towards the north, with sub-basins to the east and west separated by an area where the succession is thinner. The post-Devonian erosion of the Awaynat Wanin Group in the north of the Ghadames Basin is also clearly displayed.

Within the Libyan part of the Ghadames Basin the Middle to Upper Devonian sedimentary rocks are generally subdivided into the Awaynat Wanin Group and the Tahara Formation (Fig. 6.1). The type section of the Awaynat Wanin Group is situated to the NE of the Gargaf Arch at the Awaynat Wanin water well. This geographic area is positioned approximately on the border between the Murzuq and Ghadames basins, and sedimentary rocks in both basins can be correlated with each other. The contact between the Awaynat Wanin Group and the Lower Devonian succession is an unconformity (Gundobin 1985; El Rweimi 1991), with the boundary marked by an abrupt upward transition from sandstones to black shales.

In the Libyan Ghadames Basin the Middle to Upper Devonian succession has been subdivided into a variable number of cycles by numerous authors, the variability probably stemming from lateral and regional facies variations. An example of this may be in the northern part of the Ghadames Basin where the sandstones normally present in the upper part of the A.W. II Formation are absent (El-Rweimi 1991). This regional change in facies results in the formation of a shale dominated succession which contains sedimentary rocks of A.W. II, A.W. III and A.W. IV age, separated by subtle, and easily missed, unconformities or correlative conformities. A summary of the subdivisions of the Devonian succession in the Libyan Ghadames Basin is given figure 6.7 while general lithological descriptions of these formations are displayed in figure 6.8.

The highly radioactive Cues (or Tornoceras) limestones of late Frasnian age form a 3 to7 metre thick bed that can be traced throughout much of the Ghadames Basin on well logs and seismic sections (Bracaccia et al. 1991). Analysis of biostratigraphic data from the Middle to Upper Devonian succession has identified late Eifelian, early Frasnian and Famennian unconformites (Belhaj 1998) (Fig. 6.9), as well as the partial erosion of the Devonian succession from the SE of the Ghadames Basin. Palynological data from wells A1, A2, A3, A4 and A5-NC169 have also been used to identify an unconformity between sedimentary rocks of Givetian and Frasnian age (Elzaroug and Lashhab 1998) (Fig. 6.9).

Hasi (1995) interpreted the Tahara Formation succession as a coarsening and shallowing-upward sequence set containing 16 parasequences. El-Mehdawi (1998) also identified this shallowing-upward succession in the southern part of the Ghadames Basin, where proximal marine deposits are overlain by terrestrial/ fluviatile sedimentary rocks. An isopach map of the Tahara Formation within the Libyan Ghadames Basin (Fig. 6.10) illustrates the regional south to north increase in thickness of the Tahara Formation and localised thickness changes. Bracaccia et al. (1991) identified thickness and facies variations within the Devonian succession proximal to the Tripoli-Tibesti uplift and Gargaf Arch (Fig. 6.10), considering them to be active since at least early Silurian time. Thickness changes in the Lower Devonian succession in the southern part of the Ghadames Basin were also observed by Belhaj (1998), who interpreted them to be the result of a differential movement on a northerly extension of the Brak Bin Ghanimah uplift in the Ghadames basin. However, as outlined in chapters 3, 4 and 5, the ENE-WSW oriented Garaf Arch structure was not formed prior to the Devonian, although the Brak Bin Ghanimah uplift (Fig. 5.27) affected the Devonian succession in the Murzuq Basin (section 4.7).

Algerian Ghadames Basin

The Middle to Upper Devonian succession in the Algerian Ghadames Basin comprises shales, siltstones and subordinate sandstones, overlying an end Early Devonian/base Mid Devonian unconformity (RRI 1989; Ford and Muller 1995; Anadarko 1993a; 1996a; 1996b). This tectonic phase caused uplift of a number of pre-existing structures, including the Hassi-Messoud ridge and the High Plateau (RRI 1989), and the Ahara Arch (Anadarko 1996a; 1996b; L. Villa Lobos 1996, pers comm.). This phase of uplift and erosion provided the source of localised ferruginous coarse clastics, and was succeeded by a basinwide transgression during the Eifelian that covered most coarse clastic source areas (RRI 1989). During the Eifelian and Givetian a succession of shallow shelf shales and sandstones were overlain by 'deeper' shelf sandstones and shales (Anadarko 1996a). Minor tectonic activity during the early Late Devonian (Frasnian) in the Algerian Ghadames Basin (RRI 1989; Anadarko 1993a; 1996a; 1996b) uplifted several structures, eroding the pre-Frasnian sedimentary rocks from the crests of many of these structures (RRI 1989; Anadarko 1996a; 199b). During the Frasnian the localised deposition of carbonates occurred on the crests of several structural highs, while anoxic conditions prevailed in the deeper areas of the sub-basins (RRI 1989; Anadarko 1996a; 1996b). Comparable shallow shelf and deep marine facies were laid down during the Fammenian in the north and central parts of the basin (Anadarko 1996a; 1996b), while offshore transition zone, shoreface, barrier island and delta front facies were deposited in the southern part of the basin on the northern flank and crest of the Ahara Arch (Anadarko 1993). The deposition of these Fammenian sedimentary rocks is truncated in some areas by an unconformity (Anadarko 1993a; 1996b), while the Fammenian and the upper part of the Frasnian are completely eroded from other areas (Anadarko 1993; 1996a). A phase of tectonism during the uppermost Devonian is also inferred by RRI (1989) and Anadarko (1996a; 1996b), resulting in the uplift and erosion of Frasnian and

Fammenian sedimentary rocks from localised areas within the basin. RRI (1989) also inferred an increase in the coarse clastic influx to the basin during this time, predominantly from the south, although subordinate amounts were derived from the High Plateau to the north. Figure 6.5 gives a summary of the unconformities in the Algerian Ghadames Basin during the Devonian that are inferred by RRI (1989); Anadarko (1993a; 1994; 1996a; 1996b) and Ford and Muller (1995).

6.2.3. *Illizi Basin: Early Devonian stratigraphic evolution.*

A summary of the main subdivisions of the Devonian succession in the Illizi Basin can be seen in figure 6.1. The Tamelrik and Oued Samene formations of Gedinian to Siegenian age that crop out on the southern flank of the Illizi Basin were studied in detail by Dubois et al. (1969) and Bennacef et al. (1971) (Fig. 6.11). Emme (1988) subdivides the F6 reservoir sequence in the southern Illizi Basin into the C1, C2 and C3 sub-units of Early Devonian age (Fig. 6.1). These three cycles are usually separated by ferruginous crusts, with the lower two cycles deposited in a braidplain and the uppermost cycle containing marine fossils (Dubois et al. 1969). This interpretation contrasts with that of Bennacef et al. (1971) who identified five cycles within this succession.

The fluvial character of the Lower Devonian succession in this region was noted by Bennacef et al. (1971) and Emme (1988), the latter author interpreting the Siegenian (F6 to C3) sandstones as alluvial plain, point bar deposits. A *Tigillites* ichnofabric was also observed in the F6 to C2 sandstones (Oued Tifernine Formation) by Emme (1988). In the NW of the Illizi Basin the Lower Devonian F6 to C2 sandstones and shales were interpreted as tidal channel and offshore bar deposits by Alem and Benebouche (1995). They also inferred that the absence of the uppermost C3 sand body in this area resulted from post depositional erosion.

The Tamelrik and Oued Samene formations are gradationally overlain by the Emsian Orsine Formation (figures 6.1 and 6.11) which is coeval with the Ouan Kasa Formation of the Murzuq Basin. The Orsine Formation, also known as the F4 and F5 sandstones (Fig. 6.11), was studied on the southern flank of the Illizi Basin by Emme

(1988), who inferred that these sandstones and siltstones were deposited in marginal marine and shoreface environments during a regressive cycle. Marine incursions during deposition of the Lower Devonian Tamelrik and Oued Samene formations are also interpreted by Bennacef et al. (1971).

The ichnofabric in the Lower Devonian Oued Tifernine Formation (F6-C2) is inferred to be good evidence of at least one major marine incursion during the Lower Devonian, as well as the relative sea level rise associated with deposition of the upper part of the Oued Tifernine and Orsine formations.

The pinch out of the Lower Devonian Oued Tifernine, Tamelrik and Oued Samene formations towards the NE of the Illizi Basin is inferred by Dubois et al. (1969) to be gradational, with all of the formations present in reduced thicknesses towards the NE part of the Tihemboka Arch (Fig. 6.11). This relationship is similar to that observed on the western margin of the Murzuq Basin (Fig. 3.32), a relationship that was interpreted to be the result of a long lived decrease in accommodation space (Fig. 5.15). The data and interpretations of Dubois et al. (1969) and Bennacef et al. (1971) suggest an area within which the rate of creation of accommodation space was reduced, in this case controlled by subsidence, was present in the NE of the Illizi Basin. Dubois et al. (1969) also identified localised Early Devonian palaeohighs in the southern Illizi Basin that caused thickness changes and the noticeable rotation of palaeocurrents away from the predominant northwesterly direction. Palaeocurrent data from the F6 to C3 sandstones of Emme (1988), and the Oued Karkai Formation of Bennacef et al. (1971), also indicate meandering and anastomosing fluvial systems, migrating towards the NW (Fig. 6.12). A phase of pre-Early Devonian tectonism in southern Algeria was recognised by Bennecef et al. (1971); Emme (1988); and Anadarko (1993a), where Lower Devonian sandstones unconformably overlie the Precambrian basement. Those authors also suggested that the erosion of pre-Devonian sedimentary rocks in this region may have been a source of Lower Devonian clastic material. On the southern flank of the Illizi Basin variations in the sand/shale ratio and formation thickness of the Late Silurian to Early Devonian succession across the Fadnoun fault indicate a phase of movement on this structure during this time (Beuf et al. 1971; Emme 1988; and Anadarko 1993a).

6.2.4. *Illizi Basin: Mid to Late Devonian stratigraphic evolution.*

In the eastern part of the Illizi Basin, Middle Devonian (Givetian-Eifelian- F3) sandstones are interpreted as tidally influenced littoral bars elongated in a NW-SE direction with fine-grained inter-bar areas (Chaouchi and Kichou 1995). The orientation of these sandstone bodies is interpreted to be perpendicular to coeval palaeoshorelines, and therefore parallel to the direction of the basin floor slope, fluvial/deltaic drainage and possibly the direction of tidal action. The onlap of these Middle Devonian sedimentary rocks from south to north onto the Ahara Arch is evidence of an end Early Devonian to base Mid Devonian phase of uplift on this structure (L. Villa Lobos 1996, pers comm.).

A succession of Upper Devonian shales overlain by two coarsening-upward siltstone to sandstone cycles were observed by Emme (1988) on the southern flank of the Illizi Basin and placed within the F2 sandstones of Frasnian and Fammenian age (Fig. 6.1). The sandstones contain brachiopods, plant remains, an ichnofabric comprising *Skolithos*, *Areniculites* and *Diplocraterion* and are interpreted as low energy tidal flat deposits, deposited during two regressive cycles (Emme 1988). These sedimentary rocks are coeval with S4 and S5 on the southern flank of the Gargaf Arch, with the lower shales, of Frasnian age (Emme 1988) coeval with those observed in the Libyan Ghadames basin (section 6.2). Interestingly Emme (1988) also identified 2 to 10 centimetre thick bentonitic clay interbeds in the F2 siltstones and sandstones indicating volcanism during the upper part of the Late Devonian. Although these ash deposits can be sourced hundreds or thousands of kilometres away, the amount present most likely indicates that they were sourced relatively locally within the northern margin of Gondwana (J. Thorogood 1998, pers comm.). A possible explanation of the source of this volcanism is discussed in sections 6.5 and 6.6.

The juxtaposition of sedimentary rocks of Upper Devonian (F2) and Lower Devonian (F6) age across the Fadnoun Fault in the southern part of the Illizi Basin

(Emme 1988) indicates a phase of Late or post-Devonian movement on this structure (Emme 1988; Nigel Press Associates 1991), as well a late Silurian to Early Devonian phase of movement (Emme 1988; Beuf et al. 1971). The timing of the youngest phase of movement cannot easily be constrained in the absence of post-Devonian sedimentary rocks in this region.

6.2.5. *The Taoudini, Tindouf, Timimoun, Ahnet, Reggane, Bechar and Sbaa basins: Early to Late Devonian stratigraphic evolution.*

In the Ahnet and Reggane basins the Lower Devonian sedimentary rocks comprise bioturbated sandstones deposited on a shallow, oxygenated shelf which was stable thoughout much of the Devonian (Logan 1995). Core and outcrop data from the Reggane Basin were used by Daoudi (1995) to infer a succession of Geddinian to Siegenian sandstones as an open marine sandstone ridge complex. These sedimentary rocks are overlain by transgressive bioclastic limestones and shales of Emsian age (Daoudi 1995). In the Bechar Basin the Lower Devonian (Siegenian) deposits are interpreted by Briak et al. (1995) as deltaic to shallow marine sandy facies.

In central Morocco a series of conical Lower Devonian mudstone mounds and ridges crop out within the eastern Anti-Atlas mountain range. These mudstone mounds formed during the Emsian in deep water along a series of radial and tangential faults which were formed as a result of doming above an underlying sub-volcanic laccolithic intrusion, and are underlain by Lockkovian (Lower Devonian) basaltic, glassy lavas that erupted onto the sea floor (Belka 1997). Geochemical data from the mudstone mounds indicates the close proximity of sub-aqueous volcanic vents, with the input of thermogenic methane from underlying basaltic intrusions (Belka 1997). A similar origin is proposed for the Lower Devonian (Early Givetian) mudstone mounds and ridges that are present in the Ahnet Basin of Algeria according to Kaufman et al. (1997). These mudstone mounds are interpreted to have been deposited in deep water euphotic conditions along a series of faults that probably acted as conduits to hydrothermal fluids.

On the northern flank of the Tindouf Basin Upper Devonian and Lower Carboniferous sedimentary rocks are interpreted as fluvial, deltaic, shoreface, tidal flat and shelf facies (Vos 1977). These units are dated Fammenian to Tournaisian and are therefore coeval with the Awaynat Wanin IV and Ashkidah/Tahara formations (S4, S5, and S6) in the Murzuq and Ghadames basins (section 6.6).

During the Early, Mid and Late Devonian ferruginous oolites were deposited in a variety of forms in the Tindouf and Ahnet basins in marginal to open marine settings (Guerrak 1989; 1991). These facies occur as local ironstone deposits (LOID type), their formation and distribution controlled by localised palaeoenvironmental factors including their proximity to uplifted areas in very shallow seas (Guerrak 1991). As outlined in sections 1.3 and 1.4.2, the palaeoclimate of the North African part of Gondwana during the Devonian may have been cool or warm. However, Guerrak (1991) interpreted that the Devonian ferruginous oolite belt of North Africa formed during non-glacial conditions. In contrast, the thick succession of Frasnian siltstones deposited in Mauritania are interpreted as loess deposits, resulting from glacial erosion in South America and central Africa (Caputo 1985) (section 6.6).

Isopachs constructed for Devonian facies of the Tindouf Basin (Fig. 6.13) show the existence of a south to north directed sediment wedge, indicating the basin slope during this time. The Middle and Upper Devonian isopachs from the Tindouf Basin also show an overall thickening from south to north with the up-dip pinchout of the succession in the Taoudenni Basin and further south (Fig. 6.13). The post-Devonian erosion of Devonian sedimentary rocks from the Anti-Atlas region is also clearly shown in all of the isopachs maps displayed in figure 6.13. The Devonian ferruginous oolites in the Tindouf Basin are interfingered with marine, and rarely continental, siliciclastic sedimentary rocks, often in coarsening-upward cycles that are capped by ferruginous crusts interpreted by Guerrak (1989) to indicate periods of emergence. The Late Devonian to Early Carboniferous succession on the northern flank of the Tindouf Basin contains a number of regressive-transgressive episodes, exemplified in figure 6.14.

Geurrak (1989) identified a series of tectonic episodes during the Devonian in the Tindouf Basin. The Early Devonian phase was possibly responsible for the basaltic volcanism recorded by Belka (1997) and observed further north in Morocco. Guerrak (1989) also inferred that these tectonic events were responsible for the juxtaposition of Lower Devonian and Ordovician sedimentary rocks, erosion of Silurian facies and thickness variations of the Lower Devonian succession in the Tindouf Basin.

6.2.6. *The Kufra Basin (Libya): Early to Late Devonian stratigraphic evolution.*

Devonian facies are present in the Kufra Basin in eastern Libya (Fig. 1.2), cropping out on the SE, SW and northern margins of the basin, and also in other areas of NE Africa and the Middle East. Previous work in the Kufra Basin, Egypt, and Sudan has been hindered by the remoteness of these areas and the limited hydrocarbon exploration in these regions, while the Devonian is often absent from the Middle East or is poorly studied.

Kufra Basin. Lower Devonian stratigraphic evolution.

On the SE margin of the Kufra Basin sedimentary rocks in the Lower Devonian Tadrart Formation comprise poorly sorted, very coarse-grained to conglomeratic fluvial sandstones (Bellini and Massa 1980). Turner (1987; 1998) and Turner and Benton (1983) also interpreted these Lower Devonian sedimentary rocks as braidplain deposits, but, they also observed a thin succession of bioturbated sandstones in the middle part of the succession which are feasibly products of a marine incursion (Fig. 6.15). It was also proposed by Turner (1987; 1998) that bioturbated facies in the uppermost part of the Lower Devonian succession in this area were deposited in fluvial systems influenced by marine processes. The fluvial deposits were further interpreted by Turner (1998) to have been deposited within a Platte type braided river, similar to that interpreted for the Lower Devonian Tadrart Formation in the Murzuq Basin (section 3.2.2). Palaeocurrent data from Lower Devonian sedimentary rocks on the SE margin indicate that the primary drainage direction was towards the NW (Turner and Benton 1983; Turner 1987) (Fig. 6.15).

The lower contact of the Tadrart Formation with the underlying Akakus Formation is sharp according to Turner (1998), comprising hard, ferruginous sandstones which are often overlain by conglomerates (Bellini and Massa 1980; Turner 1987). The Lower Devonian succession is absent from the southeastern part of the Kufra Basin to the west of Jabal Arknu (Bellini and Massa 1980). On this SE margin the Tadrart Formation is overlain by the claystones and siltstones of the Ouan Kasa Formation (Bellini and Massa 1980).

On the SW margin of the Kufra Basin the Lower Devonian succession comprises the Tadrart and Ouan Kasa formations (Bellini and Massa 1980). The Lower Devonian Tadrart and Ouan Kasa formations unconformably overlie the Silurian Akakus Formation and are poorly subdivided. Bellini and Massa 1980 describe the medium to coarse-grained sandstones and subordinate conglomerates and siltstones of the Lower Devonian succession in this region as continental deposits (Fig. 6.16).

On the NE margin of the Kufra Basin the Lower Devonian Tadrart and Ouan Kasa formations comprise cross-bedded sandstones with subordinate inter-beds of siltstones and ironstones (Bellini and Massa 1980) (Fig. 6.16). In the absence of other descriptive data these sedimentary rocks are also tentatively interpreted as continental deposits.

A correlation panel of Devonian sedimentary rocks on the SE, SW and NE margins of the Kufra Basin can be seen in figure 6.16. The outcrop data illustrated in figure 6.16 closely correspond to the well data from wells A1 and B1-NC43 in the central part of the Kufra Basin where the Lower Devonian succession comprises sandstones, siltstones and rare interbeds of shale (Khattab 1984).

Kufra Basin. Middle to Upper Devonian stratigraphic evolution.

On the SE margin of the Kufra Basin the Middle to Upper Devonian Binem Formation comprises shales, silts and sandstones, with certain horizons containing plant remains and an ichnofabric comprising *Zoophycos* and *Tigillites* (Bellini and Massa 1980; Turner 1998) (Fig. 6.16). These authors interpret the Binem Formation as dominantly marine deposits, while Bellini and Massa (1980) also identify the presence of a number of continental ferruginous claystones, siltstones, and sandstones in the succession. The Middle to Upper Devonian sedimentary rocks in this region are interpreted as tidal channel, sandy tidal flats and lower shoreface deposits (Fig. 6.15) (Turner and Benton 1983; Turner 1987; Turner 1998). Uppermost Devonian sedimentary rocks from the SE of the Kufra Basin also contain evidence of a marine regression with continental and transitional facies (Bellini and Massa 1980). These authors also recognise an area west of Jabal Arknu where the Binem Formation unconformably overlies sedimentary rocks of the Akakus Formation (Upper Silurian). According to Turner (1998) the Middle to Upper Devonian succession on the SE margin of the Kufra Basin contains evidence of a number of transgressive/regressive events, controlled by variations in subsidence rates or sediment supply.

On the SW margin of the Kufra Basin the Middle to Upper Devonian Binem Formation is present south of 22°15'N according to Bellini and Massa (1980). Those authors observed that the contact with the underlying Lower Devonian facies is sharp and marked by thin beds of dark, hard siltstone containing a ferruginous cement and ironstone nodules (Fig. 6.16). Bellini and Massa (1980) divided the Binem Formation into three sub-units, each containing an ichnofabric of variable diversity and abundance. The sedimentary structures and ichnofacies observed in the Binem Formation by Bellini and Massa (1980) indicate deposition in marginal to open marine conditions for at least some of the time.

On the NE margin of the Kufra Basin the Middle to Upper Devonian Binem Formation is of increased thickness relative to the southern areas according to Bellini and Massa (1980). These authors also identify increased amounts of claystones in the lower part of the Binem Formation and also an increase in the diversity and abundance of the ichnofabric and the presence of brachiopods and gastropods (Fig. 6.16). In the central part of the Kufra Basin the Middle to Upper Devonian Binem Formation comprises alternating shales, siltstones and sandstones, with the shales between 800 and 650 feet thick (Khattab (1984).

6.2.7. *Chad, Egypt and Sudan: Early to Late Devonian stratigraphic evolution.*

In NW Sudan the Lower Devonian Tadrart Formation was studied by Klitzsch and Wycisk (1987); Klitzsch and Squyres (1990); and Klitzsch (1998) who observed medium to coarse-grained quartz sands, usually in sets of tabular cross-bedding, with thinly bedded conglomeratic sandstones. The Lower Devonian sedimentary rocks in Egypt and Sudan are interpreted as the deposits of a non-cyclic Platte type braidplain draining towards the north or locally towards the NW (Klitzsch 1998; Klitzsch and Wycisk 1987). In NW Egypt a succession of upper Emsian to lower Givetian sedimentary rocks is present (Klitzsch and Wycisk 1987 and references therein).

The Middle to Upper Devonian succession in Egypt and Sudan are marine influenced (Klitzsch 1998; Klitzsch and Squyres 1990), corresponding to those facies observed in the SE of the Kufra Basin. However, the palaeogeographic and palaeoenvironmental reconstructions of Klitzsch and Wycisk (1987) place northern Sudan and much of Egypt up-dip of any marine transgressions during the Devonian, with the region subject to continental deposition or erosion (Fig. 6.17). No data are available on the Middle and Upper Devonian sedimentary rocks in northern Sudan.

In the northern part of Chad the uppermost Upper Devonian deposits contain evidence of an early transition to continental conditions, as well as a ferruginous nodule rich horizon formed during a pre-Carboniferous hiatus (Bellini and Massa 1980).

According to outcrop and subsurface data, sedimentary rocks of Devonian age are absent from much of the western desert of Egypt, all of the eastern desert of Egypt and Sinai (Klitzsch and Squyres 1990) and much of the NW Arabian peninsular (Futyan, 1995). In part of Saudi Arabia upper Silurian strata are overlain by Middle Devonian sedimentary rocks according to Vaslet (1998).

Some of the Upper Devonian sedimentary rocks that crop out in northern Niger have been interpreted to be of glacial origin (Caputo and Crowell 1985, and references therein), although the Late Devonian age is poorly constrained. There are also reports of Fammenian glaciomarine shales in the Accra Basin of Ghana (Fig.

6.18) and Upper Devonian diamictites and other glacial sedimentary rocks in southern Africa (Fig. 6.18) (Caputo and Crowell 1985, and references therein).

6.3. The distribution and character of Devonian facies in the South American cratonic basins of Gondwana.

As outlined in section 1.3, during the Early Devonian plate tectonic reconstructions position the south pole in south/central South America, moving into South Africa during the Mid to Late Devonian. The Lower Devonian successions in the South American cratonic basins contain evidence of a transgressive episode during the Emsian, resulting in the juxtaposition of these and Wenlockian sedimentary rocks (Mid Silurian), across an unconformity representing a gap of about 40 Myr (Caputo 1985; Harland et al. 1989).

In the Acre, Solimoes, Amazonas, Parnaiba and Parana cratonic basins in the northern part of South America, (Fig. 6.18) glacigenic rocks were deposited during the Fammenian (Late Devonian) (Caputo 1985; and Caputo and Crowell 1985). There is no record of glaciation in these regions during the Early Devonian, with the earliest evidence of glaciation in South America during the Late Devonian when the south pole moved from a marine to continental position in South America (Caputo 1985). The Fammenian worldwide extinction is attributed to the Late Devonian glaciation in South America by Caputo (1985) and Caputo and Crowell (1985), who interpret a phase of regression and a disruption in oceanic circulation during this time.

6.4. A summary of the sedimentary and stratigraphic evolution of Laurentia, Baltica, Avalonia and the other main continental elements during the Devonian.

During the Devonian, while Gondwana was situated in low to mid latitudes, Laurentia and Baltica (Euramerica) moved south and collided, or moved into contact with the northern margin of Gondwana (Fig. 1.10 and 1.11) (Scotese and McKerrow 1990).

During the Devonian, the British Isles, situated in the southern part of Euramerica, were subject to fluvial and open marine conditions, as well as tectonism and igneous activity (Anderton et al. 1983; Kelly and Sadler 1995; McKie and Garden 1996). Sedimentary rocks deposited during this time are present in a number of sequences, often forming coarsening-upward cycles (Anderton et al. 1983). A succession of Middle to Upper Devonian evaporites, carbonates, mudstones, siltstones and sandstones, interpreted as aeolian and alluvial deposits with rare marine intervals occur in the East Orkney Basin (Marshall et al. 1996). In the Munster Basin in SW Ireland, Middle to Upper Devonian sedimentary rocks were deposited in terminal alluvial fans in a closed depositional system according to Kelly and Sadler (1995). A succession of uppermost Middle Devonian (Givetian) carbonates, interpreted as pelagic micrites, are observed in southern France and north Cornwall (U.K.) according to House (1995). These pelagic micrites contain a number of micro-rhythms that can be correlated between the two areas (House 1995), the possible origins of which are discussed in section 6.6.4.

During the Early Devonian the Iberian micro-plate was docked on the northern margin of Gondwana on the southern limit of the Iapetus ocean (Scotese et al. 1990). According to Keller (1997) the Lower Devonian succession in the Cantabrian region of northern Spain predominantly comprises carbonates, deposited in a hot and dry climate. A number of relative sea level fluctuations are interpreted by Keller (1997) in these carbonates, including a regionally traceable one at the base of the Lower Devonian succession (section 6.6).

In the north central North Sea, U.K, sedimentary rocks in the Clair Group are dated Mid Devonian to Early Carboniferous, comprising fluvial/deltaic sandstones with subordinate siltstones and mudstones (Mckie and Garden 1996). These deposits contain three orders of stratigraphic cycle, the origins of which are multi-variate (Mckie and Garden 1996).

In the Euramerican continent a succession of continental and mixed marine siliciclastics and carbonates of Devonian age are present (Johnson et al. 1985; Lawrence and Williams 1987; Brett and Baird 1996; Day et al. 1996) in a number of sub-basins. These sedimentary rocks were deposited in a number of coarsening and fining-upward cycles that contain, and are bound by, unconformities, traceable across a wide geographic area (Johnson et al. 1985; Brett and Baird 1996; Day et al. 1996). Lawrence and Williams (1987) interpret the palaeoenvironment in Euramerica during the Early Devonian as tropical with highly seasonal rainfall while Rust (1984) and Day et al. (1996) interpret the palaeoclimate in Canada and western Euramerica during the Mid to Late Devonian as tropical and tropical to subtropical respectively.

The increased distribution of certain Middle to Upper Devonian fishes in the east of Gondwana and the Asian terraines may be partly due to a change in global palaeogeography during this time (Burrett et al. 1990), possibly related to a global (Eustatic) rise in relative sea level during the late Frasnian. Another explanation for this increase in faunal distribution may be the collision between Euramerica and Western Gondwana (Burrett et al. 1990), removing deep oceanic barriers for faunal migration.

6.5. An overview of the sedimentary and tectonic processes in Gondwana and coeval depositonal areas during the Devonian.

The sedimentological and tectonic data and interpretations relating to Devonian facies outlined in sections 6.2 to 6.4 can be correlated with those from the Murzuq Basin detailed in chapters 3, 4 and 5. These data are summarised below, creating a regional model for the palaeogeographic and tectonic evolution during the Devonian. Although a number of the tectonic events recognised in North Africa are broadly coeval with orogenic events interpreted in other parts of the world, such as the Caledonian or Acadian orogenies, they are not discussed in these terms due to their substantial geographic displacement from the epicentres of these tectonic events and their poorly constrained driving mechanisms.

6.5.1. *Lower Devonian sedimentation and tectonism.*

As outlined above and within chapters 3 and 5, during the majority of the Early Devonian large areas of northern Gondwana, specifically Egypt, Chad, Niger and the southern part of Libya and Algeria, were covered by coarse-grained braided and meandering alluvial systems draining from the SE to the NW. In down-dip areas further north and west in northern Libya and Algeria, Tunisia and Morocco, marginal to shallow marine siliciclastic and oolitic facies pass basinward into shales and limestones. These palaeoenvironmental conditions appear to have been remarkably uniform across the North African margin of Gondwana, stretching from Egypt to the western part of Morocco.

Within localised areas of Libya and Algeria a phase of late Silurian to Early Devonian tectonism resulted in the erosion of pre-Devonian sedimentary rocks. This tectonism also formed areas within which subsidence driven, localised changes in the rate of creation of accommodation space can be recognised. An example of this differential subsidence can be observed in Lower Devonian sedimentary rocks on the western margin of the Murzuq Basin (Fig. 5.15). This phase of tectonism may also have been responsible for the formation of the 'Northern plateau' in the northern part of Morocco, Algeria and Tunisia, creating a northerly source of sediment during the lower part of the Early Devonian (Fig. 1.2). The basaltic volcanism and geothermal activity recognised in Morocco, during the Gedinnian and Emsian respectively, may indicate volcanism related to a spreading ridge, crustal extension, subduction (related to the closure of the Iapetus Ocean), or mantle convection. The limited data on the origin and nature of these igneous features, and the poorly defined plate tectonic setting of the region they occur within, make interpretations tenuous. However, the closure of the Iapetus Ocean during this time does provide a driving mechanism for this tectonism close-by.

The braided alluvial systems present in up-dip areas contain evidence of base level changes/relative sea level fluctuations (Fig. 3.33), namely changes in alluvial architecture and the interfingering of alluvial and marine influenced deposits (section 3.9.7). The overall effects of these relative sea level fluctuations were to increase absolute base level during the Late Siegenian and Emsian, resulting in the transgressive episode illustrated in figure 3.33. In the South American and North African cratonic basins a major transgression is noted during the Emsian, flooding areas exposed during previous regressive events or uplifted and eroded during the late Silurian to Early Devonian tectonic event.

6.5.2. *Middle to Upper Devonian sedimentation and tectonism.*

As outlined above and in chapters 4, 5, from the end of the Early Devonian to the early part of the Mid Devonian, a phase of tectonism occurred that resulted in the erosion of pre-Middle Devonian sedimentary rocks in the Murzuq Basin region. This phase of tectonism also resulted in the creation of and modification of sub-basins, notably in the Murzuq, Ghadames and Illizi basins. The period of relative sea level rise noted during the Early Devonian is recognised to continue into the early part of the Mid Devonian, transgressing up-dip regions and localised areas in Algeria, Libya and Saudi Arabia that were uplifted and eroded during a prior phase of tectonism. The African margin of Gondwana was subject to fluctuating continental to open marine conditions, with coarse-grained siliciclastic and fine-grained siliciclastic/oolitic deposystems dominating during the Mid and Late Devonian respectively. Similar palaeoenvironmental conditions and relative sea level fluctuations are interpreted in the other North African and Euramerican sedimentary basins while the Upper Devonian succession in central and southern Africa and the northern part of South America contains evidence of glacial activity.

During the Mid to Late Devonian the juxtaposition of facies deposited during continental to open marine palaeoenvironments is controlled by the affects of relative sea level fluctuations and localised depositional and tectonic processes. A number of the relative sea level fluctuations and unconformities identified within the Murzuq Basin can be correlated with interpretations made from facies within sedimentary basins within Gondwana and Euramerica, these are outlined in section 6.6.

During the Mid to Late Devonian Bracaccia et al. (1991) and Echikh et al. (1993) interpret a phase of rapid subsidence within the Ghadames Basin during the Middle to Upper Devonian. According to RRI (1989) this phase of increased subsidence in the Algerian Ghadames Basin occurred after a phase of tectonism during the Frasnian, the affects of which appear to be localised within the Murzuq, Ghadames and Illizi basins. This phase of tectonism was responsible for the uplift and erosion of a number of basement highs and the creation of localised areas of increased subsidence and structural highs during the Late Devonian (RRI 1989). In the Murzuq

Basin this tectonically enhanced unconformity represents the boundary between dominantly coarse-grained siliciclastic sequences and oolitic/fine-grained sequences (Fig. 4.74), while in the Ghadames and Illizi basins the unconformity is overlain by upper Frasnian radioactive shales and limestones (Fig. 6.2) and a thick succession of Fammenian sedimentary rocks. This tectonism appears to be directly responsible for a change in the amount and character of sediment supplied to the basin, probably due to the modification of alluvial drainage systems and sediment source areas. The sedimentary processes active during the Late Devonian may also have differed from those that prevailed during the Early and Mid Devonian due to glacially driven changes in regional/global climate, palaeo-oceanography and eustatic sea level.

The bentonites present in the F2 Upper Devonian sedimentary rocks in the southern part of the Illizi Basin indicate active volcanism during this time. The origin of these extrusive products is unknown but the local source within Gondwana may suggest hot spot, rift or subduction driven volcanic activity in the craton during the Frasnian or Fammenian.

In the northern part of Chad the contact between deposits of Late Devonian and lower Carboniferous age is a ferruginous nodule rich horizon formed during a pre-Carboniferous hiatus (Bellini and Massa 1980), rather than the result of any localised tectonic activity during this time. The relatively up-dip position of this region means that it is primarily an area of sedimentary by-pass and erosion except during periods of very high magnitude base level rise.

6.6. The regional and global correlation of relative sea level fluctuations during the Devonian and their possible driving mechanisms.

Within the Devonian successions outlined in sections 6.2 to 6.5 a number of relative sea level fluctuations can be interpreted, compared and correlated with those from the Devonian succession in the Murzuq Basin. The correlation of these relative sea level fluctuations can be problematic due to the variable quality or absence of biostratigraphic data within the datasets.

As outlined in section 1.2 the contact between the Silurian and Lower Devonian successions in Libya is generally recognised as an unconformity. In the Kufra Basin this surface is further defined as a type 1 sequence boundary that can be traced throughout the Palaeozoic basins of Libya, formed during a basinward shift in facies, associated with erosion of the exposed shelf and underlying sediments (Turner 1998).

The sequence and parasequence set boundaries interpreted in the Devonian succession in Libya are ferruginous-stained and pitted (sections 3.9 and 4.7). Surfaces with a similar character are observed in the Tindouf Basin within coeval ferruginous oolites facies by Guerrak (1989) and may have a similar sequence stratigraphic significance.

6.6.1. *Relative sea level fluctuations during the Early Devonian.*

Many of the relative sea level fluctuations interpreted within the Early Devonian succession can be traced across the North African margin of Gondwana, some of which can be correlated with similar events in coeval sedimentary basins worldwide. As such these relative sea level fluctuations are interpreted to be of eustatic origin, with a variable amplitude and frequency, modified by localised tectonic and depositional processes. The correlation of the individual Lower Devonian sequences interpreted within the Murzuq Basin with coeval sequences worldwide is problematic due to the variable quality of the sedimentological and biostratigraphic data. Therefore, any sequence stratigraphic interpretations of these data need to be carefully evaluated.

Sedimentary rocks of Lower Devonian age from a number of areas, outlined in chapters 3 and 5 and sections 6.2 to 6.5, have been interpreted in a sequence stratigraphic framework (figures 3.32, 3.33, 4.71and 4.76). These interpretations have been compared and correlated with previous studies on coeval successions, including work by Johnson et al. (1985); Keller (1997); and Turner (1998).

The pre-Lower Devonian unconformity interpreted within the Murzuq Basin can be traced throughout much of North Africa indicating a widespread regression that is often coeval with tectonism. As outlined in chapters 3 and 5, in the Murzuq Basin the Lower Devonian succession comprises six sequences, the juxtaposition of which indicate an overall rise in relative sea level during this time (Fig. 3.33). This Early Devonian transgression is also interpreted in Euramerica by Johnson et al. (1985), in the Algerian and Libyan parts of the Ghadames Basin by Bekkouche (1992) and Massa (1988) respectively, in the Bechar Basin by Briak et al. (1995) and in Spain by Keller (1997). This rise in relative sea level during the early part of the Devonian continues throughout the Early Devonian and increases in magnitude, resulting in a major transgression by the late part of the Epoch. The effects of this transgression can be recognised throughout the Murzuq, Bechar (Briak et al. 1995), Ghadames (Bekkouche 1992; Poyntz 1995), Illizi (Emme 1988) and Reggane (Daoudi 1995) basins.

On the southern flank of the Illizi Basin a number of cycles are identified by Dubois et al. (1969) and Bennacef et al. (1971) in the Lower Devonian succession which are correlated with those interpreted in coeval facies on the SW margin of the Murzuq Basin in figure 6.19. According to Turner (1998) in the Kufra Basin the Lower Devonian succession can be subdivided into two, the lower unit comprising a sequence containing Lowstand and Transgressive Systems Tract deposits, with the upper unit comprising Lowstand Systems Tract deposits in the lower part of a type 1 sequence (Fig. 6.20). Briak et al. (1995) also states that the Lower Devonian succession in the Bechar Basin comprises Lowstand Systems Tract deposits overlain by Transgressive Systems Tract deposits.

The marine influenced Transgressive Systems Tract facies interpreted by Turner (1998) in the middle part of the Lower Devonian succession in the Kufra Basin occur at a broadly comparable level to those interpreted by Dubois et al. (1969) and Bennacef (1971) on the southern flank of the Illizi Basin and are interpreted as coeval (figures 6.11 and 6.19). The absence of such well constrained marine facies in the Murzuq Basin may be simply due to the relatively up-dip position of this area when compared to the Illizi and Kufra basins. The accurate correlation of this period of relative sea level rise between the Murzuq, Illizi and Kufra basins is tentative due to the lack of detailed biostratigraphic data and the poorly constrained effects that localised depositional processes may have had in these areas. However, it is proposed that this period of relative sea level rise may correspond to the base of S3 in the Murzuq Basin where the sandstone rich braidplain deposits S2 are abruptly overlain by the interbedded flood plain sandstones and siltstones present in S3.

6.6.2. *The origin of the relative sea level fluctuations during the Early Devonian.*

The relative sea level fluctuations inferred for the Murzuq Basin outlined in sections 3.9; 6.6.1 and figure 3.33 are considered to have different amplitudes and frequencies (Fig. 6.21), and therefore may be driven by a variety of mechanisms (Fig. 2.6). Across the entire present day North African margin an unconformity is recognised between sedimentary rocks of Silurian and Early Devonian age (Turner 1998), although in the Murzuq Basin and immediately surrounding area, this unconformity is enhanced by localised tectonism (section 1.2.3). The widespread continuity and character of this basal Devonian unconformity is interpreted to result from a eustatic fall in sea level (figures 6.21 and 6.22) modified by the aforementioned tectonism. In the Murzuq Basin and immediately surrounding region the timing of this eustatic fall in relative sea level is also complicated by post depositional erosion and the progradational nature of the underlying Upper Silurian facies (section 1.2.3). Bellini and Massa (1980) suggest that in the northern part of the Libyan Ghadames Basin this fall in relative sea level occurred at the end of the late Silurian. The origin of the tectonic event within the Libyan and Algerian sedimentary basins near the Silurian/Devonian boundary is difficult to establish. However, the phase of basic intrusive and extrusive volcanism noted in late Silurian and Early Devonian facies in Morocco may be associated with a mid-ocean ridge, mantle plume, or given the plate tectonic setting outlined in section 1.3, subduction related volcanism. Whatever the origin of this volcanism, it may be this event resulted in the transferral of intraplate stresses into the continent, activating structures in the Murzuq Basin area. Other possible causes for this tectonic event may be, (1) a deformation 'event' centred south of the Murzuq Basin, evidence for which has been subsequently removed, or (2)

Mantle convection processes resulting in localised tectonism within the 'stable' craton.

The rise in relative sea level inferred in the Murzuq Basin in this study during the lower part of the Early Devonian (Fig. 6.21), and worldwide by other workers (Fig. 6.22), is recognised to be of global (eustatic) significance. This eustatic rise in sea level occurred near the base of the Early Devonian and continued throughout the Early Devonian Epoch, affecting the Murzuq, Ghadames, Illizi, Kufra and Bechar basins. This period of eustatic sea level rise resulted in an upward shift in the graded profile in continental areas, and an increase in marine influence and deepening of palaeowater depth in marginal to open marine settings down-dip. If this eustatic cycle terminated at the end of the Early Devonian it would cover a maximum of 21 Myr as suggested by data from the Murzuq Basin and surrounding region. Here the Lower Devonian succession is separated from the Middle to Upper Devonian succession by a pronounced unconformity that is detailed in sections 5.4.2, 5.5.1 and 5.5.2. The quantitative eustatic curve of Johnson et al. (1985) from Euramerica (Fig. 6.23) does not show a pronounced fall in eustatic sea level at this stratigaphic level, rather that the period of relative sea level rise that started in the Early Devonian continues into the Givetian Stage of the Mid Devonian. The continuation of this transgressive event into the Mid Devonian is also recognised in the Algerian Ghadames Basin by Bekkouche (1992) and in the Kufra Basin by Turner (1998). Re-evaluation of the relative sea level curve constructed from Devonian facies in the Murzuq Basin also highlights the continuation of this cycle into lower part of the Mid Devonian (Givetian). During the mid-Givetian a period of relative sea level fall terminates the Early Devonian sea level cycle, rapidly succeeded by a rise in relative sea level at the base of the following cycle (Fig. 6.21). In the Murzuq Basin this rise in relative sea level during the mid-Givetian is interpreted to occur at the base of PS4, and is recognised as a major flooding event (Fig. 4.76). Therefore the eustatic sea level rise that was initiated in the lower part of the Early Devonian (early Gedinnian) continues into the upper part of the Mid Devonian (mid Givetian) for approximately 31 Myr, the duration of which corresponds to a 2nd order eustatic cycle of Vail (1977). The duration and stratigraphic position of the relative sea level cycle within the Murzuq Basin and surrounding region closely matches that of the eustatic cycle I of Johnson et al. (1985) (Fig. 6.23) confirming its eustatic origin. Sea level cycles of 2nd order duration are believed to result from continental scale changes in dynamic topography (Miall 1997).

Lower Devonian facies outcropping in the Murzuq Basin, and published data from other areas, reveal a number of higher frequency, lower magnitude relative sea level fluctuations contained within the 2nd order eustatic cycle (Fig. 6.21). The precise duration of these higher frequency cycles cannot be ascertained due to limited biostratigraphic data but are most likely to result from cycles of 3rd or 4th order duration. As there is no record of glaciation anywhere in the world during the Early Devonian according to Caputo (1985) and Miall (1996), it is unlikely that any of the relative sea level fluctuations during this time were of glacioeustatic origin. In the absence of glacioeustasy, these sea level cycles of 3rd and 4th order duration can result from orbital cyclicity, climatic cycles, regional basement loading and the transferral of intraplate stress (Miall 1997).

These high frequency relative sea level fluctuations act to modify the overall rise in sea level associated with the 2nd order cycle (Fig. 6.21), resulting in the changes in alluvial architecture and the juxtaposition of continental and marine facies that are interpreted in Lower Devonian succession in the Murzuq Basin.

Bekkouche (1992) interprets nine members in the Lower Devonian Tadrart Formation and seven in the Ouan Kasa Formation in the Algerian Ghadames Basin. These members cannot be correlated with those in the Murzuq Basin but they are probably of a similar duration and origin. Several of the 3rd or 4th order relative sea level fluctuations can be traced between basins, such as the mid Early Devonian (top Gedinnian) marine incursion (Fig. 6.22), and also the sequence boundary and relative sea level rise interpreted to have occurred at, or near, the base of the Emsian stage (Fig. 6.22). The latter unconformity and subsequent relative sea level rise corresponds to the contact between the Tadrart and Ouan Kasa formations in the Murzuq Basin, outlined in figures 3.32 and 3.33. This contact between the Tadrart and Ouan Kasa formations also represents a surface across which there is a marked change in facies and sediment character (Fig. 3.1). This change may be the result of the alteration of the basin profile during the formation of the preceding sequence boundary, possibly coeval with tectonism, or the removal of siliciclastic sediment source areas during the subsequent transgression.

6.6.3. *Relative sea level fluctuations during the Mid to Late Devonian.*

Sedimentary rocks of Mid to Late Devonian age from a number of areas, outlined in chapters 4 and 5 and sections 6.2 to 6.7, have been interpreted within a sequence stratigraphic framework. This sequence stratigraphic framework has been compared to and correlated with previous studies on coeval successions in widely spaced areas of North Africa and around the world which also highlight a number of similar relative sea level fluctuations (Johnson et al. 1985; House 1995; Kelly and Sadler 1995; Brett and Baird 1996; Day 1996; Day et al. 1996; Leonard 1996; Marshall et al. 1996; McKie and Garden 1996; Turner 1998), a summary of which can be seen in figure 6.22.

As outlined in chapters 4 and 5, in the Murzuq Basin the Middle to Upper Devonian succession comprises a maximum of six sequences, containing eight parasequence sets and numerous parasequences (Fig. 4.76). The Eifelian to mid Givetian succession in the Murzuq Basin is interpreted to occur in the upper part of a 2^{nd} order eustatic sea level cycle (the upper part of cycle I of Johnson et al. (1985) that began during the Early Devonian (figures 6.21, 6.22 and 6.23). This cycle terminates in the mid Givetian when a rapid rise in relative sea level is interpreted to have occurred as the base of PS4, defining the base of the successive cycle (cycle II of Johnson et al. (1985) which continues into the lowermost Carboniferous (figures 6.21, 6.22 and 6.23). The rise in relative sea level during the early part of the Mid Devonian, in the upper part of cycle I, is interpreted by Bellini and Massa (1980) and RRI (1989) to have invaded Africa as far south as the present day 15°N (Fig. 6.24). The sequence boundary associated with the eustatic fall at the end of cycle I, that can be traced throughout much of Libya and Algeria, is locally enhanced by tectonism.

Other relative sea level changes that can be traced between basins occur during the late Frasnian and late Fammenian (Fig. 6.22), the latter relative sea level fluctuation includes a regression, the affects of which can be traced in the Murzuq Basin, northward into the Libyan Ghadames Basin and southward into Chad. The regional extent of this regression can be seen in gamma ray wireline data from the Upper Devonian/Lower Carboniferous sedimentary rocks in well AA6-NC7A in the northern Ghadames Basin which closely resembles the wireline trace from coeval facies within the Murzuq Basin , i.e.: it has a regressive/transgressive u-shaped response (figure 5.3 and 5.4). These data therefore demonstrate represent the affects of a regressive/transgressive event that can be traced across much of North Africa.

6.6.4. *The origin of the relative sea level fluctuations during the Mid to Late Devonian.*

The relative sea level fluctuations inferred for the Murzuq Basin that are outlined in sections 4.9; 6.9.3 and figure 4.76 are considered to have different amplitudes and frequencies, and therefore may be driven by a variety of mechanisms (Fig. 2.6). Relative sea level fluctuations of Mid to Late Devonian age with a similar character have been inferred from coeval facies in North Africa and worldwide (figures 6.21 and 6.22). The correlation of these Mid to Late Devonian relative sea level fluctuations between several basins in Gondwana and Euramerica implies a eustatic control on their formation.

As outlined in sections sections 6.6.2 and 6.6.3 the relative sea level fluctuations inferred in the lower and middle part of the Mid Devonian occur in upper part of a 2nd order eustatic cycle (cycle I of Johnson et al. 1985) (figures 6.22 and 6.23). The abrupt mid Givetian fall and subsequent rise in relative sea level interpreted in the Murzuq Basin and throughout the Ghadames and Illizi Basins corresponds with the boundary between eustatic cycles I and II of Johnson et al. (1985) (Fig. 6.23). The post Givetian Devonian succession is interpreted to occur during the eustatic cycle II of Johnson et al. (1985) which continues into the Early Carboniferous, containing a number of higher frequency/lower magnitude relative sea level fluctuations. The eustatic sea level curve derived from Middle to Upper Devonian sedimentary rocks in the Murzuq Basin is a composite curve comprising 2nd, 3rd, 4th and possibly 5th and 6th order cycles Fig. 6.21). The 3rd to 6th order sea level cycles can result from a variety of driving mechanisms including orbital forcing, regional plate kinematics, the transferral of intraplate stress, and glacioeustasy; the latter factor is believed to be of particular importance during the Late Devonian. These interpretations correspond to those of Bracaccia (1991) who interpreted the Middle to Upper Devonian sequences of the Ghadames basin as 2nd order sedimentary cycles, while House (1995) recognises high frequency Milankovitch cycles in coeval facies. In contrast the long term sedimentary cycles identified in Upper Devonian sediments in the Munster Basin of Ireland are interpreted as sedimentary cycles derived from subsidence and/or sediment flux in the basin (Kelly and Sadler 1995). The 2nd order Mid to Late Devonian eustatic cycle is inferred to be the result of continental scale changes in dynamic topography (Miall 1997).

The relative sea level curve and the position of some key sequence stratigraphic surfaces interpreted from sedimentary rocks in the Murzuq Basin (Fig. 6.21) are compared to similar data from coeval facies in Euramerica in figure 6.22. This figure shows the correlation of early and mid Givetian, late Frasnian and late Fammenian sequence boundaries, and periods of relative sea level rise, between geographically displaced basins. As described in section 4.9 the Middle to Upper Devonian succession in the Murzuq Basin is interpreted to have been deposited during a series of TST, HST and FSST's, with no evidence of abundant LST deposits. In the Kufra Basin, Turner (1998) interpreted that the Middle to Upper Devonian succession comprises a substantial thickness of TST deposits, forming a series of backstepping retrogradational parasequence sets capped by a major flooding surface (Fig. 6.20). Although the precise age of these units is poorly constrained the maximum flooding surface recognised by Turner (1998) in the upper part of the Middle/Upper Devonian succession in the Kufra Basin (Fig. 6.20) may correspond to the mid-Givetian or early Fammenian period of maximum relative sea level rise inferred in the Murzuq Basin (Fig. 5.29). The relative sea level fluctuations during the upper Frasnian and

Fammenian may be associated with the glacial events recognised during this time, recording changes in ice and oceanic water volume.

In the Murzuq Basin the tectonically enhanced unconformity formed during the Frasnian is interpreted as a surface across which there is a change in basin physiography and depositional processes. These changes affected the lithological response of depositional systems to subsequent relative sea level fluctuations, such as the character and timing of the formation of parasequences, parasequence sets, sequences and the key surfaces within these stratigraphic subdivisions.

Chapter 7. Conclusions and recommendations for future work.

A variety of outcrop data from the Murzuq Basin have been detailed and interpreted within chapters 3 and 4, and are correlated with subsurface data and their interpretations from the basin within chapter 5. These interpretations describe aspects of the sedimentological evolution of the Murzuq Basin during the Devonian which are compared to and correlated with interpretations from coeval sediments in other areas of North Africa and around the world in chapter 6. A number of key points have been raised within these sections and are discussed below, followed by a brief recommendation for future work on related research projects.

7.1 Basin profile characteristics of the North African margin of Gondwana during the Devonian.

The sedimentary and tectonic evolution of the North African margin of Gondwana during the Palaeozoic was strongly influenced by the character of the basin profile, itself the product of the underlying crust. As detailed in section 1.4.1 within the Murzuq Basin this basin profile sloped at a low angle towards the present day NW although localised modification of this slope occurred during the Palaeozoic as a result of the multiple tectonic events outlined in sections 1.2.1, 1.2.3, 5.7, 5.7.1, 6.5.1 and 6.5.2. Although a detailed analysis is beyond the scope of this study, these tectonic episodes are recognised to have caused epeirogenesis as well as localised uplifts and variations in subsidence. The background rate of subsidence within the Murzuq Basin appears to have been low for much of the Palaeozoic, although as detailed in sections 5.7, 5.71 and 6.5, differential subsidence is noted to have occurred during the entire Devonian in localised areas of Libya and Algeria.

The variety of structures interpreted from outcrop and subsurface data within the Murzuq Basin, and their influence on subsidence within localised areas of the basin, indicate the variable response of the underlying basement during tectonism. The Murzuq Basin comprises a number of different areas within which there are variations in the rheological properties of the underlying basement (section 1.2). These basement heterogeneities include a number of large N, NE, and NW striking structures in the

Murzuq, Ghadames and Illizi basins that were active at different times during the Devonian. The large structural features noted within northern Gondwana were also responsible for localised uplift and erosion as well as causing differential subsidence. The mechanisms driving these phases of tectonism are poorly constrained although the transferral of intraplate stress throughout the continent is the most likely cause. The presence of extrusive volcanic products in North and West Africa suggests active magmatism due to rifting, subduction or mantle plume activity, any of which could have been responsible for the tectonism throughout much of northern Gondwana that are summarised below and in sections 5.5.1, 5.5.2, 6.8.1 and 6.8.2.

7.2. The sedimentary and tectonic evolution of the North African margin of Gondwana during the Early Devonian.

During the Early Devonian large areas of the North African margin of Gondwana, including the Murzuq Basin, were situated in up-dip/continental settings as illustrated in sections 3.2.2, to 3.2.4, 5.6.1 and in figure 7.1. Up-dip areas on the continent such as the Murzuq Basin were covered by large, coarse-grained, 'Platte' type braided alluvial systems draining from the SE to NW. In the Murzuq, Illizi and Kufra basins the siliciclastic bedload within the alluvial systems was derived from the erosional reworking of the pre-Devonian succession to the SE, while the Northern Plateau provided a localised source of sediment in northern Algeria (section 6.2.1). The phase of late Silurian to Early Devonian tectonism identified in the Murzuq Basin and surrounding region caused the uplift of much of the North African part of Gondwana. This uplift took the form of regional epeirogenesis that resulted in the erosion and peneplanation of large areas of the craton (sections 1.4.1 and 6.5.1), as well as creating localised sources of sediment and regions of increased basin slope (Fig 7.1). The phase of tectonism during the late Silurian also caused differential subsidence to occur throughout the region, the results of which were long lived, affecting the thickness of the entire Lower Devonian succession in areas close to a number of major structural features (sections 5.6.1 and 6.5.1).

As outlined in sections 1.4.2, 3.2.4 and 5.8. The Early Devonian geomorphology, palaeoclimate and the primitive evolutionary state of terrestrial plants influenced the style of alluvial sedimentation during this time, resulting in the braiding of alluvial channels. This braiding was promoted by the unconsolidated nature of the interfluve soils which was generally poorly fixed by terrestrial plants. This situation encouraged fluvial discharge rates to be high and irregular due to a very low lag time. Therefore, interfluve areas were highly prone to erosion from the lateral accretion and avulsion of fluvial channels and also directly from precipitation and other sub-aerial erosive processes.

The bedload of the Early Devonian alluvial systems in the North African part of Gondwana generally comprised fine to very coarse-grained quartz sand deposited by the migration of straight and sinuous crested 2D and 3D bedforms. The character of the sediment and the sedimentary structures within which they occur indicates that although flow rates in the channels were moderate to high, sheet floods conditions were rare (sections 3.2.2 and 3.2.3). The deposition of fine-grained sediment during the Lower Devonian in the Murzuq Basin was often confined to the fill of isolated terminal channels (section 3.3). However, at specific levels in the Lower Devonian succession increased amounts of fine-grained overbank and flood plain deposits are observed. The widespread deposition of fine-grained sediment represents a change from deposition in widespread, high energy, braided channel systems to deposition in lower energy channels or overbank areas that are detailed in sections 3.4, 3.9.2 and 3.9.3. The rare bioturbation in these dominantly fine-grained successions also indicates that at times their deposition occurred in a mixed fluvial/paralic palaeoenvironment rather than purely fluvial conditions.

The character of these Early Devonian alluvial systems also affected the nature of the down-dip marginal to shallow marine deposystems, with the large, braidplain feeding large amounts of coarse-grained siliciclastic bedload to the braid deltas down dip, as detailed in sections 3.5.2, 3.6.2 and figure 7.1. The deltas were usually fluvially dominated although reworking by marine processes occurred during times of low discharge or tidal influx (sections 3.5.2 and 3.6.2). The deltas were flanked by small siliciclastic tidal flats and shoreface complexes (sections 3.5, 3.6, 5.6.1, 6.5.1 and figure 7.1), passing down dip into the shallow marine ramp margin that stretched many hundreds of kilometres to the NW. The influence of wave action on much of this ramp margin was minor due to the dampening affect of the basin profile and intracratonic setting of the region (section 1.3 and figure 7.1). These marginal marine palaeoenvironmental conditions prevailed in the down dip areas of Gondwana such as central and northern Libya and Algeria, Tunisia, and Morocco (section 6.5.1 and figure. 7.1).

Systematic variations in alluvial architecture occur in the lower and middle parts of the Tadrart Formation in the Murzuq Basin while the juxtaposition of alluvial and paralic facies are observed in the upper part of the Tadrart Formation (sections 3.8 and 5.6.1). These factors are interpreted to result from a series of base level fluctuations during the Early Devonian, the effects of which in the Murzuq Basin are detailed in sections 3.9 and 5.6.1. Coeval base level fluctuations are also interpreted in facies in North Africa and around the world (sections 6.2 to 6.4). These base level fluctuations occur during an overall rise in relative sea level, the effects of which can be recognised throughout much of Gondwana and Laurentia indicating a eustatic origin (sections 6.6.1 and 6.6.2).

In the Murzuq Basin the late part of the Early Devonian (Emsian) represents a change in depositional style, with the sandstone dominated Tadrart Formation abruptly overlain by the predominantly fine-grained oolitic Ouan Kasa Formation (section 3.7.2). This change in sediment character resulted from an alteration of sediment supply processes to the palaeoshoreline and associated basinal areas. These changes were driven by the combined effects of a phase of tectonism during the late part of the Early Devonian and/or a regionally extensive rise in relative sea level that are described in section 3.9.6. In the Murzuq Basin, and immediately adjacent areas in Libya and Algeria, this phase of tectonism at the end of the Early Devonian caused the movement of a number of major structures that are detailed in section 5.7. This tectonism caused the uplift and erosion of a number of areas where a thick interval of the Lower Devonian succession had previously been deposited, as well as the renewed subsidence of a number of areas that were subject to low or negative accommodation space during the Early Devonian (section 5.7). In areas of Gondwana and Laurentia unaffected by this phase of end Early Devonian tectonism the transgression initiated during the lower part of the Early Devonian continued into the Mid Devonian (sections 6.6.1 and 6.6.2).

7.3. The sedimentary and tectonic evolution of the North African margin of Gondwana during the Mid to Late Devonian.

During the Mid to Late Devonian the deposition of open to shallow marine deposits occurred in the Murzuq Basin (sections 4.1 to 4.6.3 and 5.6.2), and across much of North Africa (section 6.6.2), while marginal marine to continental sedimentation prevailed in the southern part of Egypt, Libya and Algeria (Fig. 7.2). This deposition began in the Murzuq Basin during the early part of the Mid Devonian, following on from the tectonism the area experienced at the end of the Early Devonian. However, deposition was not uniform across the entire margin of Gondwana, with non-deposition and/or erosion occurring near to a number of structures that had been uplifted during the late part of the Early Devonian (Fig. 7.2; section 5.7).

The depositional systems active in the Murzuq Basin during the Mid to early part of the Late Devonian differed from those present during the Early Devonian, in that they comprised meandering, as well as braided, alluvial systems in up-dip areas, and they transported increased amounts of fine-grained material (sections 5.4.2 and 6.5.2). The increased amounts of fine-grained sediment in this stratigraphic interval may be due to a change in the character of the sediment source areas, palaeoclimatic fluctuations or tectonism which altered the basin slope profile.

During the Mid to Late Devonian palaeoshorelines on the North African margin of Gondwana were oriented approximately SE-NW or E-W and were fed by alluvial systems oriented perpendicular to this (Fig. 7.2). The modification of the basin profile near to a number of large NW striking structures, that are outlined in sections 5.7, 5.7.1 and 6.6.2 and illustrated in figure 7.2, resulted in the localised deflection of palaeoshorelines and drainage systems away from the regional NW trend. These large structures are believed to have been active during the Mid to Late Devonian, and were responsible for localised uplift and erosion as well as causing differential subsidence. The mechanisms driving these phases of tectonism are poorly constrained although continental scale transferral of intraplate stress is the most likely cause, probably related to plate collision or subduction on the NW margin of Gondwana.

By the early part of the Late Devonian a dramatic change in sedimentation style occurred within the Murzuq Basin region, with an almost complete cessation of the deposition of siliciclastic material. This early part of the Late Devonian (Frasnian) is characterised by a phase of tectonism that uplifted a number of areas in the Murzuq Basin (section 5.7.1) and other areas of Gondwana (section 6.6.2). This Frasnian tectonism caused the erosion of sedimentary rocks near to a number of structures in the North African part of Gondwana, such as the Brak Bin Ghanimah Uplift in Libya and the Ahara Arch in Algeria (sections 5.5.2 and 6.5.2). This tectonism re-oriented sediment transport pathways, altered the basin slope profile and modified the character of sediment source areas (section 5.4.2).

Following this tectonic event a period of relative sea level rise occurred during the mid to late Frasinan that can be recognised throughout the North African areas of Gondwana and within Laurentia (section 6.6.3). The transgressive event resulted in the deposition of open to shallow marine shales, silts and ferruginous oolites that are detailed in sections 3.7, 5.6.2 and 6.6.2. This Frasnian transgression also onlapped a number of localised areas within the Murzuq and Illizi basins that were uplifted during the tectonism that occurred at the end of the Early Devonian. By the Late Devonian (Fammenian) deposition within northern Gondwana was synchronous with glaciation of the continent further to the south (section 6.3).

Deposition during the Mid to Late Devonian within the Murzuq Basin, and in other areas of northern Gondwana, was affected by a number of phases of relative sea level rise and fall that are illustrated in sections 4.7, 5.6.2 and 6.6.3. These relative sea level fluctuations caused periods of erosion and led to the juxtaposition of facies deposited in different environmental and base level conditions. One such rise in relative sea level during the Mid Frasnian transgressed a number of the areas in the Murzuq Basin (section 5.6.2) and surrounding areas (section 6.6.3) that had been uplifted during the pre-Mid Devonian and/or early Late Devonian tectonic events.

7.4. The application of Sequence Stratigraphic methods to the Devonian succession on the North African margin of Gondwana:

Relative sea level fluctuations of variable amplitude and frequency are recognised in the Devonian succession throughout the North African margin of Gondwana and are detailed in sections 3.9, 4.7, 5.6.1, 5.6.2 and 6.6. The schematic effects of one such relative sea level fluctuation on the position of the palaeoshoreline and associated deposystems in Gondwana during the Devonian can be seen in 7.3.

A number of these relative sea level fluctuations have been correlated with coeval base level fluctuations in Laurentia and other depositional basins (section 6.6). The recognition of these relative sea level fluctuations in widely spaced basins and their coeval nature suggests these are **Eustatic** sea level cycles. The Devonian to Early Carboniferous succession in the Murzuq Basin has been subdivided into a number of unconformably bound sequences, parasequence sets and parasequences that are outlined in sections 3.9, 4.9, 5.6.1and 5.6.2. These sequence stratigraphic units formed as a result of relative sea level fluctuations which modified the graded profile, palaeoclimate and alluvial architecture in up-dip areas while causing the transgression and/or regression of marginal to open marine areas down-dip.

As illustrated in section 3.9, the Lower Devonian succession in the Murzuq Basin was deposited when overall relative sea level was rising, punctuated by the effects of a number of lower amplitude, higher frequency relative sea level fluctuations. These shorter term relative sea level fluctuations directly affected base level conditions as well as influencing palaeoclimate during this time. The effects of the background rise in relative sea level during the Early part of the Devonian can be seen throughout the North African sedimentary basins and also within Laurentia (section 6.6.1). This rise in relative sea level records the transgression associated with a sea level cycle of 2nd order duration that continues into the early part of the Mid

Devonian (section 6.6.2). This 2nd order cycle also contains higher frequency, lower amplitude relative sea level cycles which are of 3rd, 4th and possibly 5th and 6th order duration (sections 3.9, 4.7, 5.6.1 and 6.6.2). As well as directly affecting relative sea level/base level the higher order relative sea level cycles almost certainly caused cyclical changes in palaeoclimate, affecting the resultant depo-systems.

The five sequences identified in the alluvially dominated Lower Devonian succession within the Murzuq Basin contain sedimentary rocks with varying architecture, specifically changes in the character of bedforms, mean grain-size and channel morphology (section 3.9.6 and 3.9.7). The occurrence of discrete bioturbated horizons at laterally continuous levels in the Lower Devonian succession in the Murzuq Basin indicates that at times the up-dip migration of paralic facies occurred due to transgression. The sequence boundaries in the Lower Devonian succession comprise laterally continuous, irregular, ferruginous-stained surface across which there is a marked change in alluvial architecture, facies and the interpreted rate of creation of accommodation space. The majority of the five sequences in the alluvial Lower Devonian succession in the Murzuq Basin are difficult to correlate with coeval successions in North Africa due to limited data, therefore their regional significance is difficult to ascertain. However, the correlation of the flooding events marking the base of sequences 6 and 7 in the Murzuq Basin in coeval deposits in other areas of Libya and Algeria, indicates that the mechanisms driving their formation were of continental, if not global significance. These flooding surfaces and the sequences within which they occur therefore formed as a result of eustatic sea level cycles.

In the Murzuq Basin the Middle to Upper Devonian succession was subdivided into six sequences within which eight parasequence sets and numerous parasequences are recognised (sections 4.9 and 5.6.2). Studies of coeval deposits in other areas of Gondwana and in Laurentia also recognise sequences in the Middle to Upper Devonian succession, although the number and character of the individual sequence stratigraphic elements varies from one area to another, as detailed in section 6.6.3. In Laurentia and Gondwana the lower part of the Middle Devonian succession was deposited during the late stages of a 2nd order cycle that also spanned the Early

Devonian (sections 6.6.2 and 6.6.4). The remainder of the Middle and Upper Devonian succession was deposited during a cycle of 2nd order duration that continues into the Early Carboniferous. Both 2nd order cycles contain relative sea level fluctuations of shorter amplitude and higher frequency, the results of which are observed within the Murzuq Basin (sections 4.9, 5.6.1 and 5.6.2) and in other basins in North Africa and around the world (section 6.6.3 and 6.6.4). The two 2nd order sea level cycles that occurred during the Devonian were driven by changes in ocean volume related to changes in continental scale dynamic topography as indicated in section 2.4.1. The shorter cycles within the Early to Late Devonian succession, of 3rd, 4th and possibly 5th and 6th order duration, act to enhance or obscure the effects of the lower frequency, higher amplitude 2nd order cycles. The driving mechanisms of these higher frequency, lower amplitude cycles are difficult to ascertain (section 2.4.1), and although the affects of glacioeustasy can be discounted during the Early to Mid Devonian such mechanisms may be important during the Late Devonian (section 6.6.4).

The end Early Devonian tectonism in the Murzuq Basin and immediately surrounding areas caused localised uplift and relative sea level fall, almost certainly obscuring the development of higher frequency, lower magnitude relative sea level cycles during this time. Similarly, there may be no evidence in the geological record of a number of other base level cycles when non-deposition and/or subsequent erosion occurred. Therefore the number of cycles preserved in the Devonian succession in the Murzuq Basin may not provide a complete representation of the overall base level cyclicity during this time.

The majority of sedimentary rocks in the individual sequences and parasequence sets were deposited when the rate of creation of accommodation space was rising or high during Transgressive and Highstand Systems Tracts although subordinate deposition occurred during Falling Stage and Lowstand Systems Tracts (sections 3.9.7, 4.7, 6.6.1 and 6.6.4). As detailed in sections 1.4 and 2.4.6, during the Devonian the North African part of Gondwana generally had a low angle basin slope that subsided at a low rate. This basinal setting resulted in the duration of

Transgressive and Highstand Systems Tracts being shortened with an early transition to extended Falling Stage and Lowstand Systems Tracts. During these times of falling and low relative sea level, wide, low angle sequence boundary surfaces formed across vast areas of the interior of Gondwana (Fig. 7.3). The Falling Stage and Lowstand Systems Tract deposits associated with these periods of relative sea level fall were very thin and/or localised with a low preservation potential due to Lowstand or Transgressive erosion. During the Early Devonian these times of low or falling base level led to the deposition of the alluvial facies comprising FA2 in the Murzuq Basin (section 3.9.1), while the sandstone-filled, erosive, prograding channels in the Middle to Upper Devonian succession may have been deposited during similar accommodation space conditions (section 4.7.2). Any sedimentary rocks deposited during relative sea level Lowstands were deposited many tens or hundreds of kilometres down the regional dip to the NW in northern Libya, Tunisia and Algeria as detailed in sections 3.9.9 and 4.9. The amount of sediment supplied to these Lowstand shorelines following relative sea level fall was also greatly reduced due to the extreme extension of alluvial profiles away from sources of sediment. Therefore, the basinal setting precluded the deposition of abundant and widespread sediments deposited during Lowstand Systems Tracts from up-dip areas of Gondwana such as the Murzuq Basin during the Devonian.

Although the Devonian stratigraphic record in the North African region of Gondwana was dominated by such times of low or falling relative sea level, the deposits of these sea level lowstands are proportionally poorly represented in the stratigraphic record.

During the Devonian the low angle basin profile, and low rates of subsidence of the North African part of Gondwana also made these periods of transgression and regression very abrupt. As shown in figure 7.3 and described in section 6.6, these relative sea level fluctuations affected much of Gondwana, with periods of relative sea level rise transgressing deep into the continental interior, while relative sea level fall exposed vast areas that were previously below base level/sea level.

7.5. Recommendations for future work.

During the research for this thesis it became apparent at a very early stage that relatively little work had been undertaken on not only the Devonian succession, but the majority of the Palaeozoic succession in North Africa. The pioneering research by Dominique Massa, Eberhard Klitzsch and Adolf Seilacher on many aspects of North African geology is still important, and this author was privileged to be able to discuss aspects of their work with them. It is clear that there is much scope for further work on most aspects of the Palaeozoic strata. However, the recommendations for future work arising out of this study will be restricted to the Devonian succession within the Murzuq Basin.

This study has outlined the evolution of the Murzuq area and has compared these interpretations with those from other areas of North Africa and other areas of the world. The amount of data utilised in this study are limited relative to the sum contained within confidential, commercial and proprietary reports covering the area. More importantly, the field data utilised in this thesis represents a small percentage of the potential data that could be collected during future studies. The following section briefly outlines a number of possible avenues that future research might explore. These studies are important in terms of pure academic research in this geographical area and stratigraphical interval but also to the hydrocarbon industry when the large hydrocarbon plays in Devonian reservoirs are considered in the adjacent Illizi and Ghadames basins.

7.5.1. *The Lower Devonian succession-future study potential.*

The Lower Devonian succession crops out on the eastern and western margins of the basin, with only limited study of both areas documented. The lateral thickness changes and sequence stratigraphic framework developed for the Lower Devonian succession in this thesis need to be further refined by future studies. The Lower Devonian succession that crops out on the eastern and western margins of the basin contains evidence of a number of relative sea level fluctuations which need to be further studied to discover their regional significance and characteristics. Extensive field studies of the Lower Devonian succession on the eastern and western margins of the basin will almost certainly allow such interpretations to be made. The outcrop data detailed and interpreted in this thesis was from a number of areas on the western flank of the basin. These data collection points are spaced from a few kilometres to many tens of kilometres apart and therefore the distance between these data points needs to be reduced by further field work in the area.

The N/S oriented outcrop pattern on the eastern and western flanks of the basin allows key surfaces, thickness variations and facies changes to be traced up-dip. The outcrop pattern unfortunately does limit the lateral (E-W) study of such features in the Lower Devonian succession, a situation further complicated by the absence of the succession from the central area of the basin. This regional change in thickness of the Lower Devonian succession is one of many identified from this study, believed to mainly result from differential subsidence and subordinate post-depositional erosion. This study has proposed the location and mechanisms that may cause these thickness changes. The limited area within which subsurface data is available from means that the precise position and geometry of these areas of differential subsidence is poorly understood. This situation will persist until future hydrocarbon exploration bridges the gap between the outcrop and continuing detail from subsurface data.

The absence of the Lower Devonian succession from the central part of the basin obviously means that there is no subsurface data relating to the character of these deposits. The abundant wireline log data from coeval facies in the Ghadames and Illizi basins, specifically gamma ray wireline log data, may make future outcrop based gamma ray spectrometry studies attractive to compare and correlate with the subsurface data. Such field based gamma ray studies could be undertaken on the eastern and western flanks of the basin as the exposure is excellent.

The Lower Devonian succession on the eastern margin of the basin has not been studied in any detail using modern geological techniques such as sequence stratigraphy. This eastern margin may contain key data confirming and refining the geological model developed from data in the western/central region of the basin. The Lower Devonian outcrop pattern on the eastern flank of the basin is also cross-cut by the Brak-Bin Ghanimah Uplift, and therefore this geographical area may enable the detailed study of its movement during the Early Devonian and interaction with sedimentary processes.

7.5.2. *The Middle to Upper Devonian succession-future study potential.*

The Middle to Upper Devonian succession crops out on the on the northern, eastern and localised areas of the western margin of the basin, as well as being present in the subsurface central area of the basin. The logged sections and study areas detailed in this thesis are spaced from several to many tens of kilometres, and therefore there are large gaps between which important sedimentary and tectonic features may be present. The gaps between the data points could be addressed during further field studies in this region.

The affects of the Frasnian tectonism and the subsequent major rise in relative sea level need to be constrained in the Murzuq Basin. While the movement of the Brak Bin Ghanimah arch is well constrained during this time, the orientation, geometry and character of the structure are poorly understood. Future studies of the Brak Bin Ghanimah uplift and associated structures on the northern and eastern flanks of the basin may help to understand the mechanisms driving this phase of tectonism. This Frasnian tectonic event is noted not only in the Murzuq Basin, but its affects are recognised in adjacent areas of the Ghadames and Illizi basins, influencing the character and distribution of hydrocarbon source and reservoir rocks.

The correlation of wireline log data relating to Devonian sedimentary rocks in the central part of the Murzuq Basin could be extended via outcrop based studies, specifically spectral gamma ray studies. These gamma ray data could be directly compared to similar data from the central area of the Murzuq Basin, potentially providing a link to the adjacent Ghadames and Illizi basins where such wireline data is abundant. The outcrop based wireline data should be collected in both the Lower and Middle to Upper Devonian successions on the east, west and northern flanks of the basin, enabling an integrated sedimentological and sequence stratigraphic model to be constructed for the Devonian Period.

This study is based on limited field work coupled with limited subsurface data and represents a 2nd order approximation to the understanding of the geological evolution of the area during the Devonian. The author looks forward to future revisions based on more extensive outcrop and subsurface data-sets in the expectation that of his conclusions some will be confirmed, some will be modified and possibly some will be overturned.

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Appendix 2.1.

Sequence Stratigraphy: Key definitions.

Sequence Stratigraphy: The study of genetically related facies within a framework of chronostratigraphically significant surfaces.

Sequence: A relatively conformable, genetically related succession of strata bounded by erosional unconformities and their correlative conformities. The sequence is composed of systems tracts and is interpreted to be deposited between falling sea level inflection points on a curve of relative sea level.

Sequence boundary: A regional scale surface of subaerial erosion and emergence associated with stream rejuvenation leading to incision and a basinward shift of facies. The erosional part of sequence boundary is referred to as an erosional unconformity and the emergent part as an interfluve.

Accommodation space: The space made available for potential sediment accumulation by sea level fluctuations (eustasy), subsidence and to a lesser degree, compaction. It is therefore controlled by relative sea level change. It is independent of sediment accumulation and does not, therefore, equate to water depth. Stratal patterns and facies distributions depend, in part upon the amount of space available for the sediment and the rate of change of new space added.

R' inflection point: The steepest part of the relative sea level curve during rising sea level, coincident with the maximum rate of increase in accommodation space.

F' inflection point: The steepest part of the relative sea level curve during falling sea level, coincident with the minimum rate of increase in accommodation space and, when the rate of eustatic fall exceeds the rate of subsidence, loss of accommodation space or negative accommodation.

Parasequence: A relatively conformable, genetically related succession of beds or bedsets bounded by marine flooding surfaces and their correlative conformities.

Parasequence set: A succession of genetically related parasequences that form a distinctive stacking pattern, bounded in many cases by major flooding surfaces and their correlative surfaces.

Marine flooding surface: A surface across which there is evidence of an abrupt increase in water depth. The surface is either a non-depositional hiatus or may involve submarine erosion by shoreface retreat/ravinement processes.

Ravinement surface: An erosional flooding surface formed by the passage of the wave breaker zone landwards during transgression. It is a type of and, in many cases, a sector of a flooding surface.

Depositional system: A three-dimensional assemblage of genetically related facies and facies associations.

Depositional systems tract: A linkage of contemporaneous depositional systems. They are interpreted to be associated with a specific segment of the relative sea level curve, in some sectors, particularly the coastal-shelf sector, they are locally equivalent to parasequence sets.

Appendix 3.1.

Locality numbers for the Lower Devonian succession (Tadrart and Ouan Kasa formations).

Locality number and Grid reference.		Comments
(1)	25°48'.51N 10°30'.68E	
(2)	25°48'.42N 10°30'.62E	
(3)	25°48'.37N 10°30'.62E	Log 1
(4)	25°48'.84N 10°30'.91E	Log 2
(5)	25°48'.88N 10°30'.66E	Log 3
(6)	25°48'.38N 10°30'.90E	Log 3a
(7)	25°37'.19N 10°31'.09E	Log4
(8)	24°39'.67N 10°35'.88E	
(9)	24°40'.52N 10°36'.61E	Log 5
(10)	25°55'.82N 10°43'.40E	Log 6
(11)	25°18'.34N 10°43'.71E	Log 7
(12)	25°19'.45N 10°43'.61E	
(12)	25°39'.23N 10°33'.86E	

Appendix 4.1.

Locality numbers for parasequence sets 1 and 2 (B'ir al Qasr Formation).

Locality number and Grid reference			Comments
(21)	27°33'.37N 12°39'.63E	Type section	Log 21
(22)	27°33'.88N 13°01'.19E		Log 22
(23)	27°33'.87N 13°01'.18E		" "
(24)	27°35'.83N 12°49'.22E		Log 24
(25)	27°34'.44N 12°39'.90E		Log 23
(26)	27°32'.89N 13°12'.96E		
(27)	27°32'.79N 13°19'.35E		
(28)	27°31'.12N 13°34'.33E		1.5km NNE of site left.
(29)	27°36'.72N 13°34'.87E		
(30)	27°33'.11N 13°34'.53E		
(31)	27°35'.94N 13°32'.73N		Log 26
(32)	27°33'.46N 13°33'.94E		Log 27
(33)	27°29'.71N 13°08'.51E		
(34)	27°36'.69N 13°34'.84E		
(35)	27°35'.83N 12°49'.22E		
(36)	27°36'.01N 13°33'.20E		Log 25

Appendix 4.2.

Locality numbers for parasequence sets 3 and 4 (Idri Formation).

Locality number and Grid reference		Comments
(41)	27°29'.32N 13°12'.61E	Log 30
(41)	27°27'.77N 13°11'.98E	" "
(41)	27°24'.47N 13°11'.69E	" "
(42)	27°31'.12N 13°34'.53E	Type section Log 31
(42)	27°31'.00N 13°35'.10E	" " " "
(43)	27°30'.83N 13°34'.72E	500 metres lateral to the type section
(44)	27°27'.01N 13°21'.26E	Upper Idri/Quttah contact
(45)	27°29'.42N 13°27'.73E	Log 32
(45)	27°29'.07N 13°27'.90E	" "
(46)	27°31'.25N 13°45'.80E	Log 33
(47)	27°31'.04N 13°23'.69E	

Appendix 4.3.

Locality numbers for parasequence sets 5, 6, 7 and 8 (Quttah, Dabdab, Tarut and Ashkidah formations).

Quttah Formation

Locality number and Grid reference		Comments
(46)	27°31'.25N 13°45'.80E	Log 33
(48)	27°25'.63N 13°22'.70E	Log 42
(49)	27°30'.24N 13°45'.97E	Log 41
(50)	27°28'.57N 13°46'.99E	

Dabdab Formation

Locality number and Grid reference		Comments
(51)	27°48'.57N 13°46'.99E	Log 51

Tarut and Ashkidah formations

Locality number and Grid reference		Comments
(51)	27°30'.46N 13°51'.87E	